Bulletin
of the
Geological Society of America

VOLUME 50 AUGUST 1, 1939 NUMBER 8

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Subscription, $10 per year.

Publication Office: Florida Avenue and Eckington Place, Washington, D. C.

Communications for publication should be addressed to The Geological Society of America, Prof. Charles P. Berkey, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Washington, D. C., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PRESS OF JUDD & DETWEILER, INC., WASHINGTON, D. C.

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LOWER PART OF TENSLEEP FORMATION

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1199-1226, 3 PLS., 2 FIGS. AUGUST 1, 1939

PENNSYLVANIAN FORMATIONS OF CENTRAL WYOMING

BY C. C. BRANSON

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ABSTRACT

The Pennsylvanian rocks of central Wyoming are a stratigraphic unit named the Tensleep formation. The formation includes the Tensleep sandstone of Darton and the greater part of the Amsden formation. Marine fossils are fairly abundant and indicate Des Moines age for the entire formation. Most of the Tensleep consists of sandstones which are uniform in size and shape of grain and in composition. The Tensleep was deposited in a shallow part of the Des Moines sea. The Quadrant quartzite is the thin northern extension of the formation. Correlatives in eastern Wyoming are the Hartville and Casper formations and the Minnelusa sandstone, each of which may be in part younger.

INTRODUCTION

The Pennsylvanian rocks of central Wyoming consist of the Tensleep sandstone, of some little known beds above that sandstone, and of the greater part of the Amsden formation which lies beneath the Tensleep. In 1916 E. B. Branson classified the lower part of the Phosphoria formation as Pennsylvanian in age and in 1931 the writer was of the same opinion. However the writer now considers the entire Phosphoria formation to be a unit of Lower Permian age. The elasmobranch fauna on the basis of which the determination of Pennsylvanian age was made is consistent with an assignment of Lower Permian age if the range of *Deltodus mercurii* extended into the Permian.

The purpose of this paper is to record information gathered by the writer concerning the stratigraphy, fauna, and lithology of the Pennsylvanian rocks of central Wyoming, and to discuss their relationship to Pennsylvanian beds in adjacent areas. The report is based upon field work carried on by the writer during the summers of 1935 and 1936, in the course of which all important localities in Wyoming were visited.

ACKNOWLEDGMENTS

Field work in 1936 was financed by a grant from the Penrose Bequest of the Geological Society of America. At various times E. B. Branson, Carl Jones, and Joseph Peck assisted in the field. The writer is especially indebted to the late Dr. G. H. Girty and to Dr. J. S. Williams of the United States Geological Survey for permission to examine collections in their care. These collections were reidentified by the writer in order to compare them with his own collections and to modernize the nomenclature of the earlier published lists. Dr. C. O. Dunbar identified the fusulinids and advised the writer of their age significance.

STRATIGRAPHY

THE TENSLEEP SANDSTONE

The Tensleep sandstone was named by Darton (1904, p. 397) from exposures in the canyon of Tensleep Creek on the west flank of the Bighorn Mountains. The formation was more fully described in a later paper (Darton, 1906, p. 34), in which it was characterized as a sheet of sandstone,

"the predominate rock white-to buff sandstone in thick massive beds, cross-bedded,"

the upper portion often calcareous,

"and at many points there are one or two beds of a mixture of sand and lime sediments."

The sandstone is 50 to 100 feet thick on the eastern side of the range, 75 to 100 feet thick on Tensleep Creek and 300 feet thick near No Wood

on the western side of the range. The upper limit of the Tensleep sandstone is established at the disconformity at the base of the Phosphoria formation (lower part of the Embar of Darton); where the Phosphoria is absent it is indicated at the base of the Chugwater formation. The lower limit is the base of the sheet of sandstone, the top of the Amsden formation of Darton. It will be shown that the Tensleep sandstone as defined by Darton is not a consistent stratigraphic unit and that it is inseparable from the Amsden.

THE AMSDEN FORMATION

The stratigraphic problems of the Tensleep sandstone are inextricably mixed with those of the Amsden formation. Darton (1904, p. 396; 1906, p. 31-34) defined the Amsden as a series of red shales, limestones, sandstones, and cherty beds overlying the Madison limestone (Mississippian) and beneath the Tensleep sandstone. The formation was named for Amsden Creek on the east flank of the Bighorn Mountains, near Dayton. The basal part of the formation is typically composed of red shale lying disconformably upon the Madison limestone, but in places there is a basal sandstone. The middle and upper parts of the formation carry large amounts of red, blue, purple, black, and brown chert. The total thickness in the type region is 200 to 320 feet.

A Mississippian coral, *Menophyllum excavatum*, was reported from the limestone of the lower part of the Amsden formation west of Sheridan. Darton listed Girty's identifications of a fauna collected from the upper cherty beds on North Fork of Crazy Woman Creek. The writer has seen this collection (U. S. Geol. Survey No. 2461) and identifies therein *Derbya* sp., *Orthotetina* sp., *Juresania nebraskensis*, *Edmondia* sp., *Deltopecten* sp., *Pleurophorus* sp., *Euomphalus* sp., *Pleurotomaria* sp., *Bellerophon* sp., *Ameura major*. The age is certainly Pennsylvanian.

In 1918 Branson and Greger (p. 309-326) described a fauna from the lower part of the Amsden in the Wind River Mountains and showed that these beds are of Ste. Genevieve age. The associated ostracode fauna, studied by P. S. Morey (1935, p. 474-482), confirms the age determination of Branson and Greger. Amsden fossils have been collected from a number of localities by United States Geological Survey parties and have been identified by Girty. Without exception these faunas are from higher beds of the formation and are Pennsylvanian in age. The writer has recently reviewed the fauna and has separated the lower part of the Amsden as the Sacajawea formation (Branson, 1937, p. 650-660). At the type locality on Bull Lake Creek the Sacajawea is disconformably overlain by 60 feet of laminated limestone beds which are possibly of Chester age. Chester time is represented by fossiliferous rocks within

the Brazer formation of southeastern Idaho and within the Big Snowy group of central Montana. The laminated limestones are overlain with pronounced disconformity by the basal sandstone of the Pennsylvanian portion of the Amsden. (*See* Plate 1, figure 2.)

THE TENSLEEP FORMATION

The Pennsylvanian beds of central Wyoming are not divisible into stratigraphic units on evidence available at present. The Tensleep sandstone of Darton is not a satisfactory formational unit. Its lithologic boundaries are not recognizable in most places, and its fauna does not differentiate it from the beds below. In many places, as at Stucco, sandstone is not the dominate rock of the section. The writer here proposes to extend the name Tensleep to the entire Pennsylvanian sequence and to call these beds the Tensleep formation. The type locality is the same as that of the Tensleep sandstone of Darton, the lower part of the canyon of Tensleep Creek. (*See* Plate 1, figure 1.)

The upper limit of the Tensleep formation is marked by the disconformity at the base of the Phosphoria formation where the Phosphoria occurs; (*See* Plate 2, figure 1) at the base of the Chugwater formation where the Phosphoria is absent. This includes some locally occurring beds at the top of the formation above the sandstone, such as the upper dolomite at Mayoworth and the capping chert at Stucco. The lower contact is at the disconformity between the basal sandstone of the Tensleep and the laminated limestones. The basal part of the Tensleep is seldom exposed. It may be seen on Bull Lake Creek in the Wind River Mountains, on Tensleep Creek, North Fork of Crazy Woman Creek, and at Stucco in the Bighorn Mountains.

REGIONAL DISCUSSION OF THE TENSLEEP FORMATION

The type section of the Tensleep formation is that exposed in the lower part of the canyon of Tensleep Creek on the west side of the Bighorn Mountains. The following section was measured there and checked in the walls of the tributary Canyon Creek.

Section of Tensleep formation near Tensleep, Wyoming

Chugwater formation

Tensleep formation

	Feet	Inches
1. Sandstone, light buff, coarse-grained, poorly cemented, with chert concretions containing *Composita subtilita, Derbya crassa, Marginifera haydenensis, Bellerophon* sp.	4	
2. Sandstone, gray, calcareous, compact, finely laminated; numerous small lime-cemented nodules of sand	25	
3. Limestone, light buff, compact, blocky	5	6

	Feet	Inches
4. Shale, purplish, mottled	6	
5. Sandstone, light yellow, laminated, cross-bedded, nodular	57	
6. Sandstone, gray, massive, inconspicuous cross-bedding	18	
7. Dolomite, purplish, mottled, shaly with much chert, contains unidentifiable fusulinids and corals	4	5
8. Sandstone, light buff; lime-cemented sandstone nodes in places	8	
9. Sandstone, white, cross-bedded, poorly cemented	6	6
10. Sandstone, gray, fine-grained	1	4
11. Dolomite, purplish, compact; numerous small crinoid columnals	2	
12. Sandstone, buff, massive	7	
13. Dolomite, gray to pink, compact		2
14. Sandstone, greenish; angular pieces of dolomite and chert as much as 2 inches in diameter from bed below; lower contact irregular		9–14
15. Dolomite, gray, with linear red streaks; large chert concretions	2	
16. Sandstone, light buff, calcareous	6	6
17. Dolomite, pink and gray, platy		6
18. Sandstone, gray, massive; irregular lower contact		8–24
19. Dolomite, light buff; much red chert	3	
20. Dolomite, dark gray; reddish shaly partings and green sandy layer at top	1	2
21. Sandstone, white, coarse-grained	2	
22. Dolomite, gray, compact	1	
23. Sandstone, green and gray, in irregular beds	1	6
24. Limestone, light gray, nearly half sand; chert bed and shale partings	14	
25. Sandstone, green, and purplish shale	2	
26. Limestone, pinkish, compact; irregular lower contact		0– 3
27. Limestone, buff, massive, cherty; purplish shale lenses	13	3
28. Shale, reddish to purplish; stringers of buff and green sandstone; irregular lower contact	5	8
29. Limestone, light buff, compact, cherty	16	
30. Shale, dark red; green shale lenses and flat nodules of red chert; lower contact irregular		6–12
Beds of Chester (?) age		
Limestone, light yellowish, cherty, stylolitic; angular pieces of purplish limestone in shale matrix in lower part	8	4
Shale, dark red-brown with green patches		9
Shale, light green, grading into light-gray limestone; limestone masses in shale matrix in lower part	6	2
Sacajawea formation		
Shale, red, 20 feet exposed at this place	—	—
Total	232	0

The several minor disconformities within the formation have not been successfully correlated with similar disconformities in other sections.

The section exposed in the vicinity of Stucco, about 50 miles northwest of Tensleep Creek, illustrates the lateral variation in the purely lithologic formation defined by Darton as the Tensleep sandstone. A section was measured there by Brainerd and Keyte (1927, p. 173-174), who on lithologic grounds assigned only 36 feet of sandstone and an overlying chert bed to the Tensleep. The following section was measured by the writer.

Phosphoria formation

Tensleep formation

	Feet	Inches
1. Sandstone, light gray, with lenticular bed of chert which contains *Schuchertella pratteni, Dictyoclostus hermosanus, Linoproductus prattenianus, Cancrinella boonensis, Juresania nebraskensis, Marginifera haydenensis, Mesolobus mesolobus lioderma, Squamularia perplexa, Composita subtilita, Spirifer rockymontanus, Neospirifer dunbari, Ambocoelia planoconvexa, Allorisma terminale, Acanthopecten carboniferus?, Plagioglypta?* sp., *Pleurotomaria* sp., *Naticopsis* sp., *Pseudozygopleura* sp., *Soleniscus* sp., *Ameura major, Lophophyllum profundum,* fenestrellinid bryozoa, crinoid columnals	4	6
2. Sandstone, light buff, cross-bedded; thin limestone and shale beds	18	
3. Chert, purple, with *Dictyoclostus hermosanus, Linoproductus prattenianus, Mesolobus mesolobus, Spirifer rockymontanus, Composita subtilita, Derbya* sp.		8
4. Shale and sandstone, calcareous	12	
5. Sandstone, light gray	13	
6. Limestone, light gray, cherty	10	
7. Shale, sandy; thin beds of limestone and brownish sandstone	23	
8. Limestone, cherty, sandy at top	12	
9. Sandstone, cross-bedded; variegated shales	65	
10. Sandstone, pinkish and gray, cross-bedded	6	
11. Limestone, gray	1	
12. Sandstone, brownish, cross-bedded	5	
13. Limestone, brownish, nodular	2	
14. Sandstone, gray; and sandy shale	8	
15. Limestone, brownish and gray, nodular; much black chert and crystalline calcite in lower part	10	
Total	190	2

Sacajawea formation

Shale, red-brown; buck-shot iron concretions; hematite beds; brown limonitic shale	70	
Sandstone, light brownish-gray, massive	15	
Shale, red, filling irregular surface on Madison limestone	0–2	
Total	87	0

A similar section is exposed near the mouth of Shell Canyon, 30 miles to the southeast.

On the east side of the Bighorn Mountains a complete section is exposed on North Fork of Crazy Woman Creek near Buffalo. The following is a section measured by the writer.

Chugwater formation

Tensleep formation	Feet	Inches
1. Sandstone, pinkish, fine-grained, cross-bedded	35	
2. Dolomite, pink	2	8
3. Sandstone, gray, cross-bedded	8	3
4. Dolomite, pink, weathers in caverns	1	1
5. Sandstone, light yellowish, calcareous, cross-bedded	17	6
6. Dolomite, pinkish, compact	9	
7. Sandstone, gray, cross-bedded	12	
8. Dolomite, yellowish, compact	12	
9. Sandstone, pinkish, fine-grained, cross-bedded	36	
10. Sandstone, yellowish, dolomitic		5
11. Dolomite, light yellowish, with small cavities where fossils have been dissolved		8
12. Dolomite, pinkish, grading downward into sandy dolomite	2	10
13. Sandstone, pinkish, dolomitic	31	
14. Dolomite, gray, sandy, cherty at top	4	4
15. Sandstone, buff to pinkish, cross-bedded	45	
16. Dolomite, pinkish		7
17. Chert, with drusy quartz		5
18. Sandstone, yellowish, dolomitic	1	2
19. Shale, purplish, sandy		6
20. Dolomite, purplish, compact		11
21. Sandstone, yellowish, lower contact irregular	8	
22. Dolomite, pinkish	6	
23. Sandstone, pink, dolomitic	4	
24. Dolomite, gray to pinkish, sandy, brecciated at top	4	3
25. Limestone, gray, compact	2	
26. Shale, purplish, calcareous	2	
27. Sandstone, gray, fine-grained	2	
28. Limestone, gray, compact		10
29. Dolomite, buff, sandy	5	
30. Limestone, pinkish, compact, cherty	1	6
31. Sandstone, buff, calcareous	2	2
32. Dolomite, buff		5
33. Sandstone, buff, calcareous, cherty at base	2	
34. Limestone, gray, hackly	2	
35. Sandstone, maroon and gray, shaly	3	
36. Limestone, gray, compact; buff sandstone	7	3

	Feet	Inches
37. Sandstone, red and yellow, calcareous	1	5
38. Limestone, gray, compact; brecciated chert	3	
39. Limestone, gray; purplish shale		11
40. Sandstone, pinkish, cross-bedded	6	
41. Limestone, gray, compact	7	
42. Chert in limestone matrix	5	
43. Limestone, gray, hackly	9	
44. Chert, with drusy surfaces, in limestone	10	
45. Limestone, buff, hackly, cherty	42	
46. Shale, red, nodular		10
47. Limestone, pinkish-gray, cherty	19	
Total	376	11

Sacajawea formation (?)

Shale, red	18	
Sandstone, buff, cross-bedded	105	

Madison limestone

The section is unusual in that there is much dolomite in the upper part.

The upper part of the Tensleep is exposed on the dip slope of the east face of the Bighorn Mountains near Mayoworth post-office, about 30 miles south of the previous locality. At the top above the massive sandstone is 22 feet of cherty dolomite beneath the Chugwater red shales. The limestone and chert of the highest bed contain *Lophophyllum profundum, Fenestrellina* sp., crinoid columnals, *Orbiculoidea* sp., *Derbya crassa, Rhipidomella carbonaria, Meekella striatocostata, Dictyoclostus hermosanus, Marginifera haydenensis,* and *Wellerella* sp. Twenty miles south of Mayoworth, near Bar C Ranch, the massive sandstone is overlain by green shale and by a peculiar conglomerate which contains limestone pebbles and chert and quartz fragments. The conglomerate carries numerous indeterminate pelecypods and gastropods. A similar conglomerate occurs at the base of the Chugwater near No Wood post-office on the opposite side of the range.

In the Owl Creek Mountains the Tensleep crops out beneath the Phosphoria, which makes the dip slope on the north side of the range. The only complete section exposed is in the walls of the canyon of the Bighorn River about 7 miles south of Thermopolis. The section measured by the writer follows:

Phosphoria formation, lower part poorly exposed

Tensleep formation

	Feet	Inches
1. Sandstone, buff, calcareous, cherty	1	
2. Sandstone, gray, finely laminated, cross-bedded	10	

	Feet	Inches
3. Sandstone, gray, sugary, contorted cross-bedding	22	4
4. Sandstone, calcareous, cross-bedded	35	
5. Dolomite; sandy and dolomitic sandstone; solution caves extending upward into overlying bed	7	6
6. Sandstone, light gray, cross-bedded	100±	
7. Dolomite, gray, compact, crystalline, with chert seam at top and yellow sandy shaly layer at base. Shaly layer contains conodonts of the genera *Idiognathodus, Streptognathodus, Prioniodus,* and *Hindeodella,* elasmobranch shagreen granules, *Helodus* sp., *Cladodus* sp., and silicified specimens of minute gastropods, ostracodes, foraminifera, *Fusulina* n. sp. and *Staffella* sp.	6	3
8. Sandstone, buff and gray, cross-bedded	50	
9. Dolomite, gray, compact, cherty	3	9
10. Sandstone, buff, coarse-grained, massive, cross-bedded	22	
11. Dolomite, gray, compact	4	
12. Unexposed, probably limestone	10	
13. Sandstone, gray, coarse-grained, thin-bedded	5	
14. Unexposed	2	
15. Limestone, gray, cherty	8	
16. Sandstone, gray, coarse-grained, cherty at top	9	
17. Limestone, light gray, crystalline, cherty	2	
18. Sandstone, light buff, cross-bedded	20	6
19. Limestone, gray, sandy, compact, cherty near top and with very sandy layer near top	25	
20. Limestone, gray, dense, cherty	11	
21. Limestone, gray, compact; green shale stringers; much solution work	5	6
22. Limestone, gray, compact, cherty	18	
23. Shale, green; mottled greenish and buff limestone at top	12	
24. Limestone, gray, compact, cherty at top	5	
25. Shale, green; thin limestones	7	
26. Limestone, gray, compact, crystalline, cherty	4	5
27. Limestone, light gray and light buff, compact; green sandstone partings	3	
28. Limestone, light gray, compact, cherty	6	
29. Shale, green, platy		2
30. Limestone, buff, crystalline, glauconitic; unidentifiable fossils	3	3
31. Shale, light green, sandy		7
32. Limestone, light buff, compact	6	5
33. Unexposed	2	
34. Limestone, light gray to buff, compact, shaly in lower part, stylolitic	9	6
35. Limestone, reddish-buff, crystalline	6	6
Total	443	8

The underlying beds, probably Sacajawea, are seen at the surface only as red clay soil.

In the Wind River Mountains two fine sections are exposed; one in the canyon of Middle Fork of Popo Agie River, the other along Bull Lake Creek 40 miles to the northwest. The section on Middle Fork of Popo Agie River (*See* Plate 3, figure 1) is as follows:

Phosphoria formation

Tensleep formation	Feet	Inches
1. Sandstone, dark gray, dense, impure, cross-bedded; chert layers	3	6
2. Sandstone, light gray, cherty, thin-bedded	4	6
3. Sandstone, light gray, massive, cherty at top	16	
4. Sandstone, light gray, fine-grained, poorly cemented, cross-bedded, contorted cross-bedding in some beds	54	
5. Sandstone, white, fine-grained, poorly cemented	13	
6. Sandstone, gray to brownish, finely laminated, cross-bedded	9	4
7. Sandstone, light gray, fine-grained	6	
8. Sandstone, light gray, fine-grained, cross-bedded	17	
9. Sandstone, light gray, laminated, poorly cemented, cross-bedded, extremely contorted cross-bedding near top	18	
10. Sandstone, light gray, fine-grained, cross-bedded	32	
11. Dolomite, light buff, compact; numerous cavities left by solution of fossils; solution channels, caverns, stylolites; lacks bedding	25	
12. Sandstone, greenish-gray; chert nodules containing *Dictyoclostus hermosanus, Linoproductus prattenianus, Squamularia perplexa, Composita subtilita*	8	6
13. Dolomite, light gray, compact, cherty		10
14. Sandstone, light gray, coarse-grained, cross-bedded	2	
15. Sandstone, light gray, calcareous, fossiliferous	2	
16. Sandstone, buff, laminated, cross-bedded; calcareous, solution channels	27	
17. Sandstone, light gray, calcareous; crinoid stem ossicles and fusulinids (?); sandstone stringers from overlying bed	4	
18. Sandstone, light gray, calcareous, cross-bedded; solution channels	61	
19. Sandstone, white, quartzitic	1	2
20. Sandstone, light gray, compact, cherty	11	
21. Sandstone, light gray, fine-grained, massive	6	
22. Limestone, light gray, compact	1	4
23. Limestone, light gray, sandy, compact, shows cross-bedding in some layers	5	
24. Limestone, light gray, compact, chert at top	2	
25. Sandstone, greenish, friable; nodular limestone	3	10
26. Sandstone, white, fine-grained, cross-bedded, calcareous	11	6
27. Sandstone, light gray, compact, upper part cross-bedded	8	4
28. Sandstone, greenish, friable; calcite geodes; irregular lower contact	1	4

	Feet	Inches
29. Limestone, gray, compact, massive chert bed at top	4	1
30. Sandstone, white, friable, cross-bedded; irregular lower contact	2	5
31. Sandstone, greenish to light buff, disconformable on lower bed	0 to 1	6
32. Limestone, gray, compact, with stringers into lower bed	4 5	4 to 2
33. Sandstone, light gray, compact, cherty	2	
34. Limestone, gray, sandy; large irregular chert masses	5	
35. Sandstone, white, friable, cherty; calcareous at base	16+	
Total	398	2

About 50 feet unexposed to top of Sacajawea formation. The massive dolomite (Bed Number 11) stands as a vertical cliff of columnar appearance overhung by the sandstones above. Float apparently from the upper sandstones has yielded tracks of amphibia at this locality and on Sheep Mountain farther south in the Wind River Mountains. The presence of the tracks indicates shallow water conditions during the time of deposition of the upper sandstone beds. The section on Bull Lake Creek is as follows:

Phosphoria formation

Unconformity (*see* Plate 2, figure 1.)

Tensleep formation

	Feet	Inches
1. Sandstone, light gray, calcareous, cross-bedded; solution caverns	80	
2. Sandstone, pink, massive, calcareous	9	
3. Sandstone, pinkish to gray, laminated	8	6
4. Sandstone, pinkish, laminated, cross-bedded, cherty near top	23	
5. Sandstone, pinkish, coarse-grained, calcareous	3	1
6. Sandstone, light gray to pinkish, massive, cross-bedded; irregular lower contact	70	
7. Dolomite, pink and gray, laminated, sandy		6 to 13
8. Sandstone, pinkish, quartzitic	5	
9. Sandstone, light yellow and pink, nearly half dolomite, cross-bedded	3	
10. Dolomite, pinkish, compact, dendritic, fossiliferous; *Deltodus* sp. found in float	3	2
11. Sandstone, pink, yellow, and buff layers, massive, cross-bedded	42	
12. Dolomite, pinkish, sandy	6	10
13. Sandstone, pinkish, coarse-grained, massive, contorted cross-bedding; solution cavities	25	
14. Dolomite, light yellowish-gray, compact, fossiliferous		8
15. Sandstone, buff, coarse-grained, laminated, cross-bedded	5	
16. Dolomite, pinkish	1	5
17. Sandstone, pinkish, laminated, cross-bedded; calcareous areas	10	

	Feet	Inches
18. Limestone, light gray, compact, largely composed of colonies of *Chaetetes*	1	2
19. Dolomite, mottled pinkish and light gray, sandy, pitted by solution	6	5
20. Sandstone, light gray, contorted cross-bedding; top of bed with cylindrical solution holes 1 to 2 inches in diameter	21	
21. Sandstone, light gray, coarse-grained; irregular lower contact		4 to 6
22. Limestone, light gray, blocky, sandy near top, much solution	3	8
23. Sandstone, light gray, calcareous, cross-bedded	4	10
24. Sandstone, greenish, shaly		4
25. Sandstone, light buff, massive, cross-bedded	6	1
26. Sandstone, greenish, compact, calcareous	1	10
27. Sandstone, light gray, calcareous	8	3
28. Sandstone, light gray, massive	2	10
29. Limestone, light gray, blocky		10
30. Sandstone, light gray, massive; stringers of green sandy shale	6	2
31. Limestone, light gray, blocky, compact	1	3
32. Chert, dark, banded, nodular; contains *Spirifer rockymontanus, Composita subtilita, Linoproductus prattenianus, Pharkidonotus percarinatus, Plagioglypta* ? sp. (Collection No. 6195 U. S. Geol. Survey, collected by Eliot Blackwelder in 1910 and labeled "U. Miss.? (Amsden or) Tensleep," from Dinwoody Creek, the next canyon north of Bull Lake Creek, is of identical lithology and no doubt comes from this bed. The writer identifies these fossils as: *Spirifer rockymontanus, Composita subtilita, Derbya* sp., *Linoproductus prattenianus, Cancrinella boonensis, Dictyoclostus hermosanus, Chonetes* sp., *Deltopecten* sp., *Acanthopecten carboniferus*?, *Leda* ? sp., *Allorisma* ? sp., *Pharkidonotus percarinatus, Euphemites carbonarius, Pleurotomaria* sp., a longitudinally striated scaphopod.)	1	7
33. Limestone, gray, compact; chert nodules	2	
34. Sandstone, light gray, cherty shaly in places; contains microfossils, *Climacammina* sp., *Idiognathodus* sp., an ostracode, gastropod and pelecypod steinkerns	5	6
35. Limestone, light gray, cherty	1	8
36. Sandstone, light buff	1	4
37. Limestone, light gray, blocky, chert at top; uncollectable fossils	3	4
38. Sandstone, light gray, massive, laminated, calcareous, contorted cross-bedding	6	1
39. Limestone, light gray, compact	1	8
40. Shale, green, sandy		3 to 5
41. Sandstone, light buff, shaly at top	1 2	9 to 11
42. Limestone, white, laminated, compact; limonite concretions	3	10
43. Shale, green, surrounding masses of greenish sandstone	1	3
44. Sandstone, quartzitic; much chert	3	
45. Sandstone, white, laminated	3	1
46. Limestone, sandy; and calcareous sandstone	3	2

	Feet	Inches
47. Limestone, gray, compact, thin partings	2	11
48. Shale, green		4 to 8
49. Limestone, gray, platy		6
50. Limestone, light gray, compact; hematite concretions	1	7
51. Sandstone, greenish gray, shaly, calcareous		10
52. Limestone, nodular; green shale stringers	1	4
53. Limestone, light gray compact; chert layers, chert brecciated in places	2	10
54. Limestone, light gray, shaly	2	10
55. Quartzite, greenish gray, dense	1	8
56. Limestone, cherty; and cherty sandstone alternating with green shale	28	7
57. Limestone, light gray, compact; chert nodules	7	8
58. Sandstone, light gray slabby	8	
59. Sandstone, light gray, calcareous	6	
60. Sandstone, reddish, quartzitic, coarse-grained	12	
61. Sandstone, red and brown, coarse-grained, cross-bedded	8	
62. Sandstone, light buff, cross-bedded	16	
63. Sandstone, light yellow-brown, coarse-grained, cross-bedded; pockets of conglomerate containing chert pebbles and fragments from underlying bed; contact disconformable	30	
Total	535	5
Beds of Chester ? age		
Limestone, light gray, sandy, laminated, compact; contact disconformable	60	3
Sacajawea formation		
Limestone with Ste. Genevieve fauna above sandstone breccia with sandstone and shale matrix; disconformable upon Madison limestone	48	

At no other locality is the basal disconformity of the Tensleep formation so clearly marked.

The northernmost outcrops of rocks that have been called Tensleep rather than Quadrant are at the mouth of the canyon of Shoshone River, near Cody, Wyoming. The section is as follows:

Phosphoria formation (with basal pebble conglomerate)

Tensleep formation

	Feet	Inches
1. Sandstone, light yellowish, thin-bedded	1	
2. Sandstone, brown, white and purple, shaly, limonite concretions	3	11
3. Sandstone, light gray, laminated, cross-bedded	90	
4. Shale, yellowish, with light gray limestone	1	1
5. Sandstone, light yellow, massive	1	4

	Feet	Inches
6. Limestone, white, blocky	2	
7. Sandstone, white, massive, cross-bedded	8	1
8. Sandstone, yellowish, shaly		10
9. Sandstone, light gray; possibly disconformable on lower bed	1	8 to 8
10. Limestone, light gray and light yellow; pink chert	5	10
11. Sandstone, light yellow, calcareous; large chert nodules; chert bed at top	3	8
12. Limestone, light yellow, geodal	1	8
13. Shale, yellowish, sandy; light gray limestone with large rough chert nodules	4	2
14. Sandstone, light gray, laminated	3	3
15. Limestone, light buff, crystalline and geodal in middle portion	2	10
16. Shale, light reddish-blue; contains conodonts, small pelecypods, gastropod steinkerns	2	
17. Limestone, light gray, compact		8
18. Sandstone, purplish-red, massive	1	3
19. Limestone, dark purple, sandy; many small calcite crystals; lenses of light gray limestone	1	8
20. Limestone, light purplish to gray, contains *Allorisma terminale*	1	1
21. Sandstone, yellow, geodal, alternating with light gray, cherty limestone	5	
22. Limestone, light gray	2	8
23. Sandstone, light buff	2	5
24. Limestone, light gray, cavernous	1	1
25. Sandstone, light gray calcareous	1	8
26. Sandstone, white, friable	1	7
27. Sandstone, pinkish, quartzitic		10
28. Limestone, light gray, compact; pink chert	6	
29. Sandstone, yellow, thin-bedded; shale	8	
30. Limestone, light gray, geodal; lens of reddish sandstone in middle part	9	6
31. Limestone, light buff, compact	3	
32. Shale, green	1	1
33. Sandstone, greenish, thin-bedded		5
Total	180	11

Sacajawea formation—red sandstone mottled buff at top, above red shale

The above section lacks the dolomite beds of other sections and contains more shale. The lithology is variable in a way that is comparable to that of the formation near Stucco in the Bighorn Mountains.

South of the Bighorn Mountains the Tensleep crops out in several small ranges. Near Ervay post-office in the west end of Rattlesnake Range

200 feet of sandstone are exposed at the top of the formation. Fine exposures may be seen in the canyon of the North Platte River at Alcova and in the canyon below Pathfinder Dam, 6 miles south of Alcova. The section at Alcova was measured by Bradley (1935, p. 147-165), who gives the thickness of the Tensleep as 236 feet. (*See* Plate 3, figure 2.) Some of the beds designated by Bradley as limestone are dolomite. The section below Pathfinder Dam is given by Lee (1927, p. 51-53). Four hundred twenty-nine feet of beds are referable to the Tensleep formation. From the highest limestone bed Lee lists Girty's identifications of *Allorisma terminale, Pteria longa?, Deltopecten aviculatum, Parallelodon?* sp., *Schizodus?* sp., *Pleurophorus?* sp., *Bellerophon* n. sp., *Euphemites carbonarius?,* and *Naticopsis* aff. *N. nana.* A second limestone 50 feet lower contains specimens which Girty identified as *Productus cora, P. hermosanus, Composita subtilita, Parallelodon?* sp., *Deltopecten?* sp., *Pteria* sp., *Plagioglypta?* sp., and *Bellerophon crassus?* (U. S. Geol. Survey No. 3850). The contact with the Sacajawea (?) is 80 feet below the lower fossiliferous horizon.

At Belle Springs, 15 miles north of Rawlins, 300 feet of sandstone is exposed and is overlain by 54 feet of shale and limestone which may or may not belong to the Tensleep formation. Eight miles south of this locality the writer collected the following from a pink limestone ledge a few feet above the red shale of the Sacajawea (?): *Dictyoclostus hermosanus, Linoproductus* sp., *Echinoconchus semipunctatus?, Squamularia perplexa, Schizophoria texana?, Composita subtilita, Spirifer rockymontanus, Aviculopecten* sp., *Schizodus?* sp., and an orthoceratite cephalopod. Barnett collected from a

"pinkish sandstone band 60 feet above quartzite and just above massive sandstone"

(quoted from specimen label) 3½ miles west of Rawlins. The writer has examined this collection (U. S. Geol. Survey No. 3139) and identifies *Campophyllum* sp., echinoid spine, *Orbiculoidea* sp., *Derbya* sp., *Linoproductus prattenianus, Dictyoclostus* sp., *Echinoconchus semipunctatus, Spirifer rockymontanus, Composita subtilita, Cliothyridina?* sp., *Deltopecten* sp., *Myalina* sp., *Schizodus* sp., *Astartella* sp., *Pleurophorus* sp., *Allorisma terminale, Pleurotomaria?* sp., *Ameura* sp. Lee (1927, p. 72) lists a collection from the top limestone of the "Amsden" at this locality, but since there is an unexposed interval beneath the sandstones, the stratigraphic position is uncertain. The writer has seen this collection (U. S. Geol. Survey No. 3843) and identifies *Linoproductus prattenianus, Dictyoclostus hermosanus, Echinoconchus semipunctatus?, Chonetes* sp., *Composita subtilita, Cliothyridina?* sp. These are undoubtedly Tensleep faunas.

LITHOLOGY

The Tensleep formation is dominated by sandstone. The upper part of the Tensleep is in most places a series of massive, cross-bedded sandstones. Many of the beds are cemented by carbonates and surface leaching may have removed much cement near the outcrop. The sandstones characteristically develop small nodes in relief on weathered surfaces. These nodes are roughly spherical masses of lime-cemented sandstone which stand out when the surrounding poorly cemented grains are washed away. In most places they are about the size of a pea, but locally are an inch in diameter. Similar nodes are abundant in sandstone of the Chugwater and Sacajawea formations, and are therefore not peculiar to the Tensleep. A bed of sandstone on Bull Lake Creek (Bed No. 20 of the section) shows vertical cylinders of lime-cemented sandstone which weather more rapidly by solution than the rest of the rock so that vertical tubes are formed which remind one of troilite cavities in meteorites.

The sandstones of the Tensleep are remarkably regular in size and shape of grain and in composition. With the exception of a few coarse-grained beds, the grains range in size from 0.01 mm. to 1 mm. in diameter. Few grains are larger than 0.5 mm. and in most sands few less than 0.05 mm. The grains are angular and show little sign of wear. The few large grains, however, are fairly well rounded.

A. W. Quinn of Brown University has examined the heavy minerals in samples of sandstone from the Tensleep formation. He reports that these are notably scanty in proportion to the size of the samples. Unlike the other grains the heavy mineral grains are well rounded. The common minerals are zircon, tourmaline, leucoxene, and magnetite. Hornblende and garnet are rare. The heavy mineral suite shows little variation laterally or vertically within the formation. The source of supply of sand was constant over wide areas for a considerable length of time. Tests of the Sacajawea and Dinwoody sands show that these can not be readily differentiated from the Tensleep sands on the basis of heavy minerals.

The Tensleep formation contains little shale. There are several beds of green shale in the lower part on Bull Lake Creek and some varicolored shale in the lower part of the formation near Greybull and Cody. Elsewhere even shale partings are scarce.

Dolomite is abundant in massive beds in the upper part of the formation particularly in the southern part of the outcrop area. Limestone beds are largely confined to the lower part of the formation. Chert is abundant as nodules and nodular beds except within the massive cross-bedded sandstone of the upper part of the formation.

FIGURE 1. DISCONFORMABLE CONTACT OF PHOSPHORIA FORMATION
On cross-bedded sandstone of Tensleep formation. Contact indicated by black line.

FIGURE 2. MASSIVE SANDSTONE OF UPPER PART OF TENSLEEP
Base of Phosphoria forms ledge at top of cliff.

TENSLEEP FORMATION ON BULL LAKE CREEK

FIGURE 1. TENSLEEP-PHOSPHORIA CONTACT ON MIDDLE FORK OF POPO AGIE RIVER
Contact indicated by black line. Re-entrant near base of cliff is a thick dolomite bed of columnar appearance.

FIGURE 2. MASSIVE SANDSTONES OF UPPER TENSLEEP AT ALCOVA
Calcareous beds in lower part of cliff.

FIGURE 3. CONTORTED CROSS-BEDDING IN SANDSTONE MEMBER OF TENSLEEP
Near mouth of canyon of Middle Fork of Popo Agie River.

UPPER PART OF TENSLEEP FORMATION

The massive sandstone beds of the Tensleep are consistently cross-bedded. There is no indication of systematic direction of dip. Festoon cross-lamination of the type observed by Knight (1929) in the Casper formation was not observed in the Tensleep. In several beds, particularly in the Wind River and Owl Creek Mountains the cross laminations are contorted and in a few places the laminations have been bent into overturned folds. (*See* Plate 3, figure 3.)

A few of the cross-bedded layers are ripple marked. Branson and Mehl (1932, p. 383-398) have described tracks of a web-footed animal from a Tensleep sandstone bed on Sheep Mountain in the southern part of the Wind River Mountains, and have since found two additional types of track in a Tensleep sandstone bed on Middle Fork of Popo Agie River.

CORRELATION AND AGE

QUADRANT FORMATION

The Quadrant formation crops out in Yellowstone Park and in south central Montana. The formation was named by Iddings and Weed (1899, p. 34) from a section on Quadrant Mountain, although the name was used in earlier publications by those authors and by others. Scott (1935, p. 1011-1032) recently measured again the rocks on Quadrant Mountain and established the lower contact of the formation. Iddings and Weed at their type locality included at the base of the Quadrant 100 feet of rock concealed by talus, but apparently they did not recognize Mississippian rocks above the Madison and had no intention of including such rocks in the Quadrant. Except for the covered beds of the one section, they applied the name to the sandstone and quartzite beds and interbedded limestones. Scott applied the name Amsden to the beds below the Quadrant and above the Madison, and referred them to the Mississippian on the basis of fossils collected in central Montana. It is almost certain that the upper beds which Scott referred to the Amsden are Tensleep in age. He did not recognize the fact that the Mississippian Amsden is but the lower part of that formation (Sacajawea of this paper). The red and purple shales and overlying fossiliferous limestone of Scott's section are in the writer's opinion Sacajawea.

The writer measured a section of the beds of Tensleep age at Devil's Slide, near Gardiner, Montana. The base of the Phosphoria there consists of two feet of pebbly phosphatic sandstone overlain by two feet of conglomerate containing abundant specimens of *Bellerophon* sp. The Tensleep consists of 130 feet of gray and buff sandstone and quartzite with but one thin limestone bed. Wilson (1934, p. 372-373) has given a section at or near this locality. Below the Tensleep are stringers and

TABLE 1.—*Fauna of Tensleep formation by localities*

	Bull Lake Creek	Dinwoody Creek	Popo Agie River	Gros Ventre Mts.	Thermopolis	Cody	Stucco	Tensleep	Crazy Woman Cr.	Mayoworth	Rawlins	Leo	Shirley	Alcova	Casper	Boxelder Creek
Staffella sp					x											
Fusulina n. sp					x											
Climacammina magna	x				x											
Lophophyllum profundum								x		x						
Campophyllum sp											x					
Chaetetes milliporaceus	x															
Lingula sp				x												
Lingulidiscina sp				x												
Orbiculoidea sp										x	x					
Orthotetina sp									x							
Rhipidomella carbonaria										x						
Derbya crassa		?		x				x	?	x	?				?	
Schuchertella pratteni							x									
Meekella striatocostata										x						
Chonetes sp		x		x							x	x	x		x	
Mesolobus mesolobus							x									
Juresania nebraskensis				x			x		x			x				
Echinoconchus semipunctatus											x					
Dictyoclostus hermosanus		x	x	x			x			x			x	x		?
Marginifera haydenensis							x	x		x						
Linoproductus prattenianus	x	x	x	x			x		x		x	x	x	x		?
Cancrinella boonensis		x					x									
Squamularia perplexa			x	x			x				x					
Spirifer rockymontanus	x	x		x	?		x				x					
Neospirifer dunbari							x									
Ambocoelia planoconvexa					?		x									
Composita subtilita	x	x	x	x			x	x			x		x	x		x
Wellerella sp										x						
Allorisma terminale		?				x	x				x	x		x		
Pleurophorus sp				x					x	x	x			?		
Myalina sp											x					
Astartella sp											x					
Acanthopecten carboniferus		?					?									
Deltopecten sp		x							x		x	x		x		?
Pleurotomaria sp		x		x			x		x		x					
Strophostylus nanus									?							
Naticopsis sp							x		?					x		
Pseudozygopleura sp					x		x									
Soleniscus sp					x		x									
Bellerophon sp								x	x					x		
Euphemites carbonarius		x		?	?									?		
Pharkidonotus percarinatus	x	x														
Plagioglypta sp	?	?					?							?		
Ameura major							x		x		?					
Hindeodella sp					x											
Prioniodus sp					x	x										
Idiognathodus sp	x				x	x										
Streptognathodus sp					x											
Helodus sp					x											
Deltodus sp	x															

pockets of conglomerate which separate it from 40 feet of compact, light-gray limestone which overlies 90 feet of green and red shale and sandstone (Sacajawea?). A similar section is exposed one mile south of Livingston, Montana.

Girty has identified a Pennsylvanian fauna from the Snowcrest Range which includes *Derbya robusta?, Linoproductus prattenianus, Spirifer rockymontanus, Composita subtilita,* and *Hustedia mormoni* (Condit, 1919, p. 116). It has been pointed out by several authors that the Quadrant pinches out entirely in central Montana. The Quadrant together with the upper part of the beds which Scott referred to the Amsden are here regarded as the attenuated edge of the Tensleep sediments.

TENSLEEP FORMATION

The Tensleep formation crops out in the Gros Ventre Range of western Wyoming. The upper part of the formation, exposed at the mouth of the canyon of Gros Ventre River, consists of 53 feet of cross-bedded, light-gray sandstone with ripple marked layers, overlain by 15 feet of limestone. The Phosphoria has a basal conglomerate which rests disconformably upon this limestone. Beds of the Tensleep formation are the principal rocks involved in the Gros Ventre rock slide. Blackwelder (1913, p. 174-179) listed Girty's identifications of fossils from beds identified as Amsden in the Gros Ventre Range. The writer has examined these collections (U. S. Geol. Survey Nos. 6189-6193) and identifies *Lingula* sp., *Lingulidiscina* sp., *Derbya crassa, Juresania nebraskensis?, Dictyoclostus hermosanus, Linoproductus prattenianus, Chonetes granulifer, Composita subtilita, Spirifer rockymontanus, Pleurophorus* sp., *Conocardium* sp., and crinoid, echinoid, and bryozoan remains. A lower faunule is composed almost entirely of generically identifiable gastropods and pelecypods, but contains also *Composita subtilita, Squamularia perplexa,* and *Spirifer rockymontanus?.*

CASPER FORMATION

The Casper formation was defined by Darton in 1908 (p. 407, 418) and was further discussed in 1909 (p. 13). Darton suggested the name for the Carboniferous rocks of the crest of Casper Mountain and along both flanks of the Laramie Mountains. He recognized that the Casper included beds of Madison, Amsden, and Tensleep age. The Pennsylvanian portion of the Casper in Casper Mountain has been studied by Lee (1927, p. 48-51). His sections show the upper beds to consist of 215 feet of massive cross-bedded sandstone overlying 100 feet of brown quartzitic sandstone. The lower beds are not well exposed, but there is more than 100 feet of limestone and shale above the red shale of the

Sacajawea (?). Near Douglas, Lee measured more than 1000 feet of beds which he would refer to the Casper. The upper bed consists of 40 feet of quartzitic breccia, a lithology not found at other localities. Pennsylvanian fossils have been reported at several places in Casper Mountain and to the south in the Freezeout Hills.

In the Laramie Mountains the Casper formation is largely composed of massive reddish sandstone beds and pinkish stylolitic limestone beds. Hall (1852, p. 399-414) early reported *Productus semireticulatus* and *Avicula? custa* from these beds along the North Platte River above Fort Laramie. His specimens are in the collection of the United States Geological Survey and are *Dictyoclostus hermosanus, Linoproductus prattenianus,* and *Allorisma* sp. The writer collected from a stylolitic limestone near the middle of the Casper 10 miles southeast of Laramie *Meekella striatocostata, Linoproductus prattenianus,* and *Juresania nebraskensis.* The Casper formation is well exposed on the eastern front of the Laramie Mountains, notably in the quarries at Horse Creek and Altus. Lee (1927, p. 41) made a detailed section of the beds at Horse Creek. The writer obtained fine collections from several beds of this section. From the limestones with blue chert near the top of the Casper he collected *Derbya* sp., *Meekella striatocostata,* and *Dictyoclostus* sp. The thick limestone beds in the middle of the formation are very fossiliferous. From four beds in this series the writer obtained *Derbya crassa, Juresania nebraskensis, Dictyoclostus* sp., *Linoproductus prattenianus, Marginifera haydenensis, Chonetes granulifer, Ambocoelia planoconvexa, Neospirifer dunbari, Composita subtilita, Dielasma bovidens* (a broad spatulate variety). From the red sandstone near the base of the formation abundant specimens of *Linoproductus prattenianus* and a specimen of *Meekella striatocostata* were collected. Lee (1927, p. 42) lists Girty's identifications of specimens from Mill Creek, 1½ miles north of Horse Creek, and from Altus, which in addition to some of the species obtained by the writer include *Cyclotrypa barberi, Productus pertenuis (=Cancrinella boonensis), Echinoconchus semipunctatus, Allorisma terminale,* and *Bellerophon crassus.*

Lee proposed that the name Casper be abandoned and replaced by extension of the name Ingleside, a formation with type section in Colorado, for the upper beds, by the name Fountain for the lower Pennsylvanian beds, and by the name Madison for the Mississippian limestones. Knight (1929, p. 49) rejected this nomenclature and restored the name Casper to the beds which Lee would call Ingleside. The writer agrees with Knight that the Ingleside is but the thin southern extension of the Casper and that the use of that name would be inappropriate. He would refer the restricted Casper of Casper Mountain to the Tensleep and he

believes that the so-called possible Fountain of Casper Mountain (Lee, 1927, p. 49) is Sacajawea.

The Casper on the east slope of the Laramie Mountains is lithologically similar to the Hartville formation, and as it is near the outcrops of that formation, it might well be called Hartville, although the higher beds are not recognized at Horse Creek. The Casper of the Laramie Basin is primarily of Tensleep age, but may include younger beds. It is lithologically intermediate between the limestones and shales of the Hartville and the predominantly sandy beds of the Tensleep. The Fountain formation is either not present in central Wyoming or is not recognizable in the lower Tensleep. Miller and Thomas (1936, p. 715-738) recently described a cephalopod fauna and listed a number of fossils from the Casper of the Laramie Basin. They conclude that the Casper may include beds of Lower and Upper Pennsylvanian time and possibly even Lower Permian. The writer is doubtful of the presence of beds younger than Des Moines.

HARTVILLE FORMATION

The Hartville formation crops out over a wide area in the Hartville uplift in eastern Wyoming. The formation was named by Smith (1903, p. 2-3) to include all the rocks below beds he identified as Opeche and Minnekahta and above the Mississipian Guernsey limestone (equivalent to Madison). The Hartville formation has been recently investigated by Condra and Reed (1935), who point out that it may include some Permian rocks. The greater part is Pennsylvanian as shown by fossil lists given by Condra and Reed and by fusulinids described by Thompson (1936, p. 95-113), who regards them as indicating that the lower part of the Hartville is Morrow in age, the middle part Middle Cherokee and the upper part Upper Pennsylvanian. Condra and Reed divided the Hartville into six divisions of which Division VI (the lowest) is regarded as probably Pennsylvanian, Division V Marmaton or older, Divisions IV and III Des Moines, Division II Missouri and Division I Pennsylvanian or Permian. The writer collected from the top of Division IV *Dictyoclostus americanus, Juresania nebraskensis, Linoproductus prattenianus, Composita subtilita, Spirifer rockymontanus, S. occidentalis, Neospirifer dunbari, Tetrataxis* sp., *Staffella* n. sp., *Fusulina* n. sp. (identical with that from the middle of the Tensleep in the Owl Creek Mountains), *F. truncatulina, Wedekindellina coloradoensis, Helodus* sp., and conodonts. From the division above in the hill along the railroad 5 miles north of Glendo he collected echinoid spines, *Meekella striatocostata, Echinoconchus semipunctatus, Dictyoclostus americanus, Linoproductus prattenianus, Juresania nebraskensis, Chonetes* sp., *Ambocoelia arcuata, Composita subtilita* and *Neospirifer dunbari.* These are Tensleep faunas.

The lower beds of the Hartville rest upon the Guernsey limestone with pronounced disconformity. A short distance below the dam on Guernsey Lake red sandstone of the basal Hartville fills a valley 50 feet deep in Guernsey limestone. In the top of the lower part of the Hartville (Divi-

TABLE 2.—*Range of species in fauna of Tensleep formation*

	Tensleep			Casper	Hartville	Minnelusa	Ingleside	Weber	McCoy	Hermosa	Wells	Cherokee	Marmaton	Missouri
	Lower	Middle	Upper											
Fusulina n. sp.	...	...	x	...	x	...	...	...	...	...	...	...	...	...
Climacammina magna	...	x	...	...	...	...	...	...	x	...	...	...	...	...
Lophophyllum profundum	...	x	x	...	...	x	...	x	...	x	...	x	x	x
Chaetetes milliporaceus	...	x	...	x	...	x	x	...	...	...	...	x	x	x
Rhipidomella carbonaria	...	...	x	...	...	...	...	...	...	...	...	...	...	x
Derbya crassa	...	...	x	x	x	x	...	x	x	x	...	x	x	x
Schuchertella pratteni	...	...	x	...	...	...	...	...	...	...	...	...	x	...
Meekella striatocostata	...	...	x	x	...	...	x	...	...	x	...	...	x	x
Mesolobus mesolobus	...	...	x	...	x	x	...	...	x	x	x	x	x	...
Juresania nebraskensis	x	...	...	x	x	x	x	x	x	x	x	x	x	x
Dictyoclostus hermosanus	...	x	x	x	x	?	...	...	...	x	...	...	...	...
Marginifera haydenensis	...	...	x	x	...	x	...	x	...	...	...	...	...	...
Linoproductus prattenianus	...	x	...	x	x	x	x	...	x	x	x	x	x	x
Cancrinella boonensis	...	x	x	...	x	...	...	...	...	...	...	x	x	x
Squamularia perplexa	...	x	x	...	x	x	x	...	x	x	?	x	x	x
Composita subtilita	...	x	x	x	x	x	x	x	x	x	x	x	x	x
Ambocoelia planoconvexa	...	...	x	x	x	x	...	...	x	x	...	x	x	x
Spirifer rockymontanus	...	x	x	...	x	x	x	x	...	x	...	...	...	...
Neospirifer dunbari	...	...	x	x	x	...	...	...	...	...	...	...	x	x
Edmondia nebraskensis	x	...	...	?	...	x	...	...	...	...	...	?	?	?
Allorisma terminale	...	x	x	?	...	...	...	...	...	x	...	x	x	x
Pharkidonotus percarinatus	...	x	...	...	...	x	...	x	...	...	...	x	?	?
Euphemites carbonarius	...	x	...	x	...	...	...	x	x	...	...	x	x	x
Ameura major	x	...	x	?	...	...	...	...	...	x	...	...	x	x
	3	12	17											
Total species in common	24			11	12	13	7	9	8	13	4	13	16	15

sion VI of Condra and Reed) there is a reddish limestone with unidentifiable fossil fragments. Condra and Reed found a disconformity at the top of this division, but believe that the greater disconformity below is more likely to be the system boundary. The writer from his examination of the lithology and stratigraphy considers it likely that these beds are the eastern equivalent of the Mississippian Sacajawea formation. Divisions II, III and IV are correlatives of the Tensleep.

MINNELUSA SANDSTONE

The Minnelusa sandstone crops out in the Black Hills of South Dakota and Wyoming. It consists of more than 400 feet of sandstone with several thick limestone beds and a basal red shale with thin limestone

beds (Darton and Paige, 1925, p. 8-9). The fossils listed by Darton and Paige did not definitely establish the age of the Minnelusa, but later collections by United States Geological Survey geologists have proved that it is Pennsylvanian. Pennsylvanian fossils have been found in upper, middle, and lower beds of the formation (U. S. Geol. Survey Nos. 4372, 4397, 5703, 5708, 5708a, 5709, 5710, 5267). Dillé (1930, p. 619-623) collected and identified from the basal red shales of the Minnelusa 20 species, of which 14 are brachiopods. Thompson (1936, p. 95-113) described three fusulinids from the lower part of the Minnelusa and stated that they indicate correlation with the lower part of the Hartville. Brady (1931, p. 183-188) measured a section of the Minnelusa near Beulah, Wyoming, and collected 20 identifiable fossil species from the upper beds. He concluded that this portion of the Minnelusa is Missouri in age. The Minnelusa certainly contains beds of Tensleep age, but may be in part younger.

The Ingleside formation of northern Colorado is regarded as a thinner portion of the Casper and as such is a Tensleep correlative. In central Colorado rocks cropping out in several regions and called locally Weber (?) and McCoy have faunas similar to that of the Tensleep.

The fauna of the Tensleep is a Des Moines fauna. The brachiopods indicate correlation with the Marmaton of the Mississippi River valley and the conodonts are definitely of Des Moines age. *Fusulina* n. sp. of the Tensleep is identical with a form in the part of the Hartville which is considered as Upper Des Moines in age. No Cherokee faunas are found in the Tensleep, but the lower part of the formation has yielded few fossils.

PALEOGEOGRAPHY

The Tensleep formation and its correlatives in Wyoming include enormous quantities of pure sandstone of fine and even grain size. The purity and physical character of the grains and the small proportion and well rounded shape of the heavy mineral grains indicate that the sand was derived from older sandstones. The greater part of the sand was deposited during retreat of the Tensleep sea. The sandstone beds of the upper part of the formation show typical shallow-water structures such as ripple marks, foot prints, and cross-bedding. During the time of maximum advance marine dolomite and limestone were deposited.

The Hartville area was the western end of the comparatively deep seaway connecting with the basin in Nebraska, Kansas, Iowa, and Missouri. The fauna is so similar in the two areas that it is certain that there was no barrier such as that postulated by Weller (1930, p. 593). The limestone beds finger southwestward into the Casper and Ingleside. Heaton (1933, p. 109-168; 1937, p. 1247-1252) has published paleogeographic

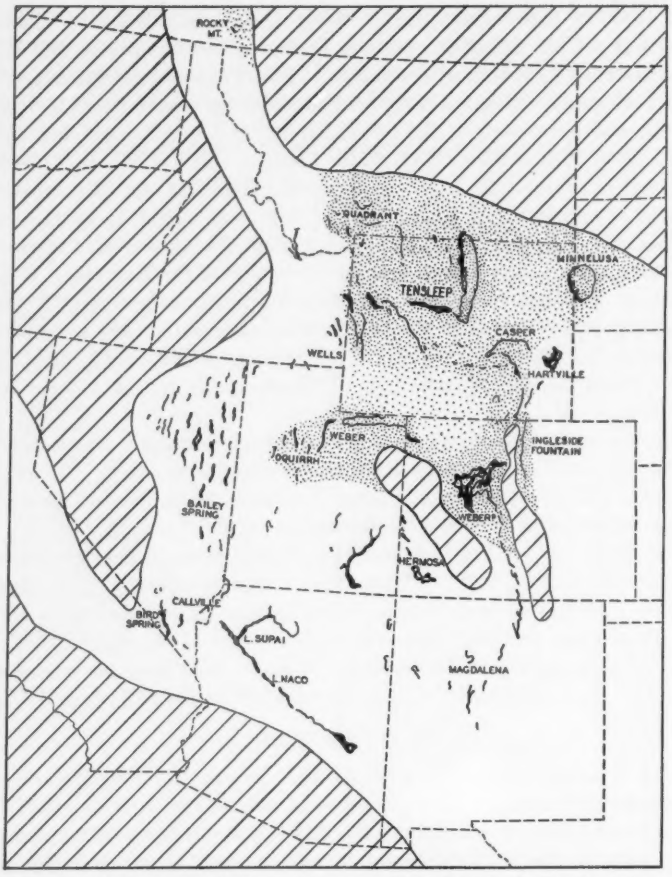

FIGURE 1.—*Paleogeographic map of western United States in Tensleep time*

Stippling shows area of sand deposition. Land areas are diagonally ruled. Present outcrops of Lower Pennsylvanian formations are shown in black with name of formation in each region (e. g. Tensleep, Casper, etc.).

maps for Lower and for Upper Pennsylvanian time in the Rocky Mountain region, but he has assumed that the higher beds of the Tensleep and of the Fountain and Ingleside are Upper Pennsylvanian. The writer in compiling the accompanying map has also necessarily assumed time equivalence that is not proved, but he has based the map upon distribution and faunal characters of formations which are at least in part of Des Moines age.

SUMMARY

The Tensleep sandstone of Darton is only locally recognizable as a lithologic unit and is not a true stratigraphic unit. No boundary can be drawn to separate it from the Amsden formation and the fauna of the

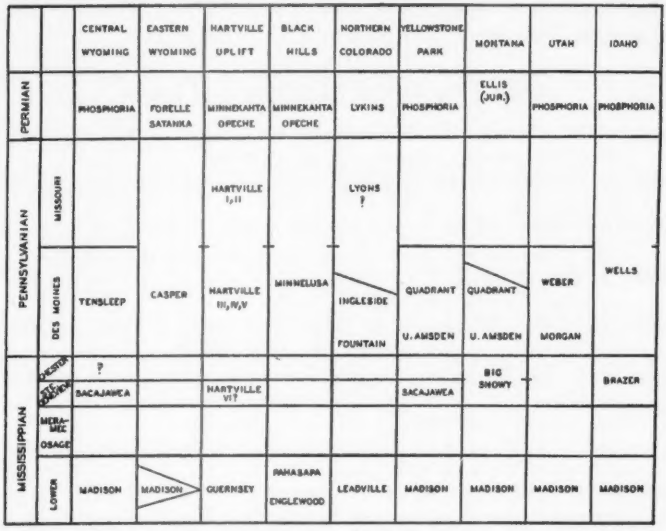

FIGURE 2.—*Correlation table of Pennsylvanian formations of Wyoming*

upper part of the Amsden is indistinguishable from that of the Tensleep sandstone. Since the name Tensleep is so well known it is here retained as the name of a formation which includes the entire Pennsylvanian sequence in central Wyoming; that is, the Tensleep sandstone of Darton, the upper part of the Amsden, and some unnamed beds above the Tensleep sandstone. No other name is available. The Amsden includes Mississippian rocks; the Casper originally included Mississippian rocks and may be in part younger than the Tensleep formation; the Quadrant is the attenuated edge of the sediments and has been applied to older and to younger beds.

The fauna of the Tensleep formation establishes its age as Des Moines and probably entirely younger than Cherokee. The Quadrant quartzite of Montana and northwestern Wyoming is, as redefined by Scott, the thin edge of the Tensleep sediments. The Casper (restricted), Hartville, and Minnelusa are in part Tensleep correlatives, but each may contain younger beds.

The high proportion of dolomite and limestone in the Tensleep formation is recognized for the first time. The sandstone beds are composed of fine quartz sands of small grain size and uniformity with extremely small heavy mineral content. The sands were derived by erosion of older sandstone beds. The thick sandstone beds of the upper part of the formation were deposited in shallow water during retreat of the sea from central Wyoming.

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Brown University, Providence, Rhode Island.
Manuscript received by the Secretary of the Society, August 10, 1936.
Project Grant 107-34.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1227-1244. AUGUST 1, 1939

ST. CROIXIAN CLASSIFICATION OF MINNESOTA

BY CLINTON R. STAUFFER, GEORGE M. SCHWARTZ, AND GEORGE A. THIEL

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ABSTRACT

Pioneer work on the St. Croixian series of Minnesota and Wisconsin was done by Owen. In Minnesota this was followed by the work of N. H. Winchell and his associates. The classification thus established has been used with but slight variations for more than 40 years. The studies of Ulrich and the more recent work of Twenhofel, Raasch, and others in Wisconsin suggest various modifications of merit which are being used more widely and which may prove very helpful in the interpretation of the St. Croixian series. It is suggested by the authors that the delimitations of the Minnesota formations be brought into approximate accord with those now being used in adjacent regions.

INTRODUCTION

This paper is presented in an effort to relate the classification of the St. Croixian series that has been inherited by the present Minnesota Geological Survey to those now used elsewhere and especially to adjust it to that commonly called the Conference Classification (Bridge, 1937, p. 234). The general relationships of certain of these classifications of the St. Croixian series are shown in Table 1, and other classifications are mentioned wherever necessary to bring out differences in usage that have contributed to the adjustments here made.

The term St. Croixian is derived from the name proposed by N. H. Winchell (1873, p. 69) for the sandstone and intercalated shales cropping out along the St. Croix River. According to his statements these attain a thickness of "about six hundred feet, in their exposures along the Mississippi River," where they occupy "a conspicuous place in the bluffs . . . below the Lower Magnesian limestone." About 20 years earlier Owen and his associates had studied this same series during a reconnaissance survey of the Upper Mississippi Valley. Owen (1852,

TABLE 1.—*Classification of the St. Croixian series*

OWEN 1852	WINCHELL 1886–1888	BERKEY 1897	WINCHELL 1905	HALL, MEINZER, and FULLER 1911	WALCOTT-ULRICH 1914	ULRICH 1924	MINNESOTA GEOL. SURV. 1932	RESSER 1933	TWENHOFEL, RAASCH, and THWAITES 1935	BRIDGE 1937	MINNESOTA GEOL. SURV. 1939
Formation 1 f (lower part)	Jordan sandstone	Jordan sandstone	Jordan sandstone	Jordan sandstone	Jordan (Winchell 1874)	Jordan sandstone	Jordan sandstone	Jordan sandstone	Trempealeau formation	Madison member	Madison (sandstone) member	Jordan sandstone	Van Oser member
Formation 1e						Trempealeau: Norwalk ss. member				Jordan member	Trempealeau formation: Jordan (ss.) member; Jordan ss. of Ulrich; Norwalk ss. of Ulrich		Norwalk member
Formation 1 d (upper part) 5th or Stillwater trilobite bed	St. Lawrence limestone	St. Lawrence dolomites and shales	St. Lawrence limestone and shale	St. Lawrence formation	St. Lawrence (Winchell 1874)	Trempealeau: Lodi shale member	St. Lawrence formation	Trempealeau formation		Lodi member	Lodi (shale and siltstone) member	St. Lawrence formation	Lodi member
						Trempealeau: St. Lawrence ls. member				St. Lawrence mem.-basal gr'sand & carg.	St. Lawrence dolomite (member)		Nicollet Creek member
Formation 1 d (main part) 3rd and 4th trilobite beds	shales					Mazomanie ss.		Mazomanie formation	Franconia formation	Bad Axe member	Franconia formation: Mazomanie ss. of Ulrich; Bad Axe (ss.) mem.	St. Lawrence formation	Bad Axe member
										Hudson member	Hudson (ss.) mem.		Hudson member
Formation 1 c (upper part)	Dresbach sandrock	Franconia sandstone	Dresbach sandstone and shale	Dresbach sandstone	Franconia (Berkey 1898)	Franconia ss.	Franconia sandstone	Franconia formation		Goodenough member	Franconia formation: Franconia ss. of Ulrich; Goodenough (ss.) mem.	St. Lawrence formation	Taylors Falls member
								Ironton formation		Ironton member	Ironton (ss.) mem.		Ironton member
Formation 1 c (main part)		Dresbach shales			Dresbach (Winchell 1888)	Dresbach ss.	Dresbach formation	Eau Claire formation	Dresbach formation	Galesville member	Galesville (sandstone) member	Dresbach formation	Galesville member
Formation 1 b - 1st and 2nd trilobite beds					Eau Claire (Ulrich MSS 1914)	Eau Claire shale				Eau Claire member	Eau Claire (sandstone) member		Eau Claire member
Formation 1 a (upper part)					Mt. Simon (Ulrich MSS 1914)	Mt. Simon ss.		Mt. Simon formation		Mt. Simon member	Mt. Simon (sandstone) member		Mt. Simon member

p. 48) called it "Formation 1" or "Lower Sandstone" to distinguish it from the sandstone known somewhat higher and cropping out at the mouth of the St. Peters or Minnesota River. It was Owen, with the help of Shumard, who first attempted to subdivide these sedimentary rocks on a faunal basis, and some of their zones are essentially the same as those that have been redefined more recently. Thus Owen's (1852, p. 52) fifth trilobite bed (*Dikelocephalus minnesotensis* zone) is the fossiliferous portion of the Lodi member; his third (*Ptychaspis* zone) and fourth (*Prosaukia* zone) trilobite beds are approximately the Hudson member, and his first trilobite bed (*Crepicephalus* zone) is a major part of the Eau Claire member. Owen, however, regarded these faunal units as zones in a single formation, a formation in fact which included the red sandstones and shales of the Lake Superior district. These latter, however, are probably Keweenawan and certainly older than the St. Croixian series.

WINCHELL CLASSIFICATION

Winchell started his intensive geological work in the Minnesota Valley where only the upper portion of the St. Croixian series crops out. This includes the St. Lawrence formation and the Jordan sandstone (Winchell, 1874, p. 147-155) which were named from localities in that region. When later his work extended to the immediate vicinity of the Mississippi River he (Winchell, 1886, p. 336-337) discovered lower beds of the series and named them "Dresbach sandrock." Although this completed the divisional names of the St. Croixian series insofar as Winchell was concerned, there remained the shales which his early table of formations shows underlying the St. Lawrence limestone and overlying the Dresbach sandstone but to which he assigned no name at the time. Later, however, Winchell (1888, p. xxii) discussed the St. Lawrence limestone and stated that

> "if the shaly beds with which it is associated, and into which it seems to graduate, be included under this term, it will include beds to the amount of nearly 200 feet".

These shales and shaly sandstones, parts of which are often dolomitic, were thereafter (Winchell, 1905, p. 268) included in the St. Lawrence. This seemed to be a natural and justifiable expansion of the original formation and was made, moreover, by the man who described it. Following this formational content, Hall and Sardeson (1895, p. 170) indicated 213 feet for the "dolomite and sandy shale" belonging to the St. Lawrence, while Norton (1897, Pl. VI) gave 230 feet for the same beds in Iowa.

As is true of the preliminary work in adjacent States, Winchell's was pioneer geology for this region, and consequently in his prelimi-

nary publications, even in Volume 1 of his final reports, there are a great many tentative conclusions and errors which had to be corrected (Sardeson, 1924). Occasionally more recent writers have referred to these tentative results of Winchell regarding St. Croixian subdivision in Minnesota but have neglected his final conclusions, thus tending to perpetuate errors which Winchell himself had corrected. Unfortunately, since Winchell started his geological work in the Minnesota Valley, where outcrops are unfavorable, he chose his type sections from incomplete exposures there, whereas the same horizons are much better shown in the bluffs along the Mississippi River. Moreover, his descriptions of the formations named are often inadequate, and the vertical extent of the rocks included in them is none too certain. Nevertheless there is sufficient outcrop of both the St. Lawrence and the Jordan at the type localities to determine the nature of these formations, and those writing on the geology of Minnesota have usually respected the priority of these names.

Berkey (1897-1898) made a detailed geological survey of the St. Croix Dalles where there are excellent outcrops of the St. Croixian series. At the village of Franconia, near the lower end of the Dalles, he found a very thick sandstone which he recognized as part of Winchell's Dresbach formation (Berkey, 1897, p. 370) or the "Basal sandstone" of Norton (1897, p. 140) lying immediately below the St. Lawrence formation. This, Berkey (1897, p. 373) states,

> "exhibits certain characters sufficiently constant to merit subdivision into at least three distinct parts. The uppermost subdivision is a sandstone exhibiting two phases: (a) an incoherent fine sand, which is underlain, (b) by more compact and thick-bedded layers. Thin seams of green shale occasionally appear in this bed. Its thickness is about 100 feet at its most favorable exposure. Fossils are not well preserved within it, and they are not abundant except in one horizon. Because of the exceptionally fine exposure of this formation in the vicinity of the small village of Franconia, this uppermost division of the series is called the Franconia sandstone."

It is thus evident that the Franconia sandstone was taken out of the Dresbach or upper part of the Basal Sandstone series and that it lies below the shales underlying the quarry rock at the type locality of the St. Lawrence. The importance of this original limitation of the formation makes it seem worth while to repeat the complete classification as given by Berkey (1897, p. 377), in the table on page 1231.

In a somewhat later section Berkey (1906, p. 233) gives a thickness of 30 to 213 feet for the St. Lawrence dolomites and shales in southern Minnesota but retains the same thickness for the Franconia sandstone. Winchell appears never to have recognized the Franconia sandstone as a distinct formation but continued to use his own three formational divisions of the St. Croixian series. Although Hall (1901, p. 168) had

recognized the Franconia sandstones and shales as lying between the Dresbach shales and the St. Lawrence dolomites and shales, Hall, Meinzer, and Fuller (1911, Pl. VI) ignored the Franconia sandstone as a Minnesota formation and returned to the usage established by

Classification given by Berkey (1897)

MAGNESIAN SERIES (Hall and Sardeson)	Shakopee dolomite New Richmond sandstone Oneota dolomite Jordan sandstone St. Lawrence dolomites and shales			
BASAL SANDSTONE SERIES (Modified from Norton)	3. Franconia sandstone (100 feet)			The St. Croix Formation (Winchell)
	2. Dresbach shales (150 feet)	*Obolella polita* zone	Greensands and shales	
		Lingulepis pinnaeformis zone	Calcareous and pyritiferous shales	
	1. The lowest formation of this series is not exposed in the Dalles area, but it includes the lowest sandstone beds and possibly also the "Hinckley sandstone," (0 to 1000 feet).			

Winchell. Ulrich (1911, pl. 27) also omitted the Franconia sandstone from his earlier classification and placed the Jordan sandstone, and doubtfully the St. Lawrence, in the middle of his Ozarkian. The balance of the series, even the Hinckley, he included in his St. Croixian.

RECENT CLASSIFICATIONS

More recently Walcott (1914, p. 354) accredited to Ulrich the classification which he used in his paper on the *Dikelocephalus and other genera of the Dikelocephalinae*. This recognized the Franconia as a distinct formation and is very important because it gives thicknesses of the formations and imposes limits not introduced in earlier discussions. In this classification the term Dresbach is restricted to the comparatively unfossiliferous sandstones above the Tostevin quarries at Dresbach, and the name Eau Claire is substituted for the highly fossiliferous beds of Winchell's type Dresbach section.

For the Minneapolis-St. Paul folio Sardeson (1916, p. 4) used a classification in which the Franconia is recognized and its thickness given. This is especially interesting because it is the first publication of the Franconia sandstone by the United States Geological Survey and in-

cludes all the major formational divisions of the St. Croixian series as used in Minnesota.

In 1919 a highly important transitional classification was used by Twenhofel and Thwaites (1919, p. 616). While this was intended as a Wisconsin classification applicable to the Tomah-Sparta region, its location is less than 30 miles east of the Minnesota State line and should be suggestive of conditions to be expected along the Mississippi River. It is especially interesting because it places the upper boundary of the Franconia at a greensand conglomerate near the base of the dolomitic layers and divides the different formations into lithological units, although it continues the use of the formational names that had been considered standard for the St. Croixian series. Twenhofel and Thwaites (1919, p. 623) regarded this as a more logical dividing plane between Franconia and St. Lawrence than that used by Hall, Sardeson, and others because they considered the greensand conglomerate as evidence of an important disconformity between members of differing lithologic content. They justified this expansion of the Franconia on the ground that the limits of Berkey's formation had not been fixed by the author.

When Ulrich's (1924) notes were finally published, about 10 years after he had given the tentative classification used by Walcott, he introduced several new names and discarded or demoted other terms long in use for the St. Croixian series. Although his classification deals specifically with Wisconsin and only in a general way with conditions in Minnesota, nevertheless it is a highly important one. Together with the preceding classification it is the real beginning of the differences in the interpretation of the beds lying between the dolomitic limestones of the St. Lawrence formation and the sandstones cropping out at Franconia. These are the sandy and dolomitic shales which Winchell had originally left unnamed but later had included in the St. Lawrence formation. Ulrich introduced the term Mazomanie to include what has turned out to be the greater part of these same intermediate beds. Although his table indicated the absence of the Mazomanie in west Wisconsin, and presumably in Minnesota, this has not been found to be the case as the fauna shows that much of that division is represented in the greensands and glauconitic sandstones that he included with the typical Franconia along the Mississippi River. This change in interpretation affects more than 100 feet of shales, glauconitic beds, and sandstones, all of which may be more or less dolomitic, that the Minnesota Geological Survey has been in the habit of including in the St. Lawrence formation as expanded by Winchell. In the same classification Ulrich (1924, p. 80) made the St. Lawrence a member of the Trempealeau, which latter he introduced as a new formation to include

the beds extending upward from the top of his Mazomanie to the top of the Norwalk or fossiliferous lower portion of the Jordan sandstone. Ulrich also limited the Dresbach to the relatively barren sandstones, above the Tostevin quarries, which have more recently been named the Galesville. The same classification is used by Ulrich and Resser (1930, p. 11) in their paper on the Cambrian of the Upper Mississippi Valley except that nothing is said about the absence of the Mazomanie in that region, and all the subdivisions of the Franconia, except the Ironton, are omitted. Also the Mt. Simon is made the basal member of the Eau Claire sandstone and shale.

Wanenmacher, Twenhofel, and Raasch (1934) made a very important faunal and sedimentary study of the Cambrian of Wisconsin. During this work there was a zoning of the St. Croixian series on a faunal basis which suggests the early works of Owen (1852) and Wooster (1882, p. 110, Pl. VI). While other workers used the faunal sequence as a basis for subdivision, the generic names of the most characteristic forms had not been definitely attached to the divisions as zonal names in published sections. The classification Wanenmacher, Twenhofel, and Raasch used transfers the Mazomanie to the Franconia, thus completing the separation of the lower beds or shales and shaly sandstone from the St. Lawrence. This tends to establish the conglomerate bed as the bottom phase of the St. Lawrence, a move which, wisely or not, has not been followed by the Minnesota Geological Survey.

Berkey's original section of the Franconia included the *Conaspis* zone only, but, on the basis of his further discussion, there is perhaps some justification for the expansion of his Franconia sandstone to include the glauconitic shales and sandstones (chiefly Mazomanie) occurring above the *Conaspis* zone. For instance, Berkey (1898, p. 272) says that "the Franconia sandstone includes the third trilobite bed of Owen" which is the *Ptychaspis* fauna. However, he follows this statement with a list of the true *Conaspis* fauna collected from the type section at Franconia and then remarks that "other localities have added" a larger number of forms which he lists. This latter is a mixed lot, chiefly from Owen's third trilobite bed and doubtless not collected by Berkey. Many of these early faunal lists from the Cambrian of the Upper Mississippi Valley need much revision because the collections from which they were compiled are labeled with such indefinite horizon names as "St. Croixian sandstone" or "Potsdam sandstone" and the true horizon of such specimens is not known with certainty. It may be noted too that the *Conaspis* zone or Taylor Falls member of the Franconia sandstone dips southward from Taylors Falls through the village of Franconia and passes beneath river level before the Osceola bridge is reached. Berkey

(1897, p. 371), however, shows about 50 or 60 feet of Franconia above river level in the Osceola section. The beds thus indicated lie much higher in the section than the sandstones along Lawrence Creek at Franconia and are wholly above the *Conaspis* zone. This might then be taken to indicate that the Franconia sandstone as a formation was intended to include a greater thickness than crops out at the type locality, but it leads to the inclusion of thick sandy shales and glauconitic beds which are not a part of the interval covered by the original description. After describing the Franconia as a rather fine-grained, quartz sandstone, Berkey (1898, p. 140-141) remarks that

> "the uniform white color varying locally to brown or yellow through ferric oxide stains, the rather angular character of the grains, the porous and friable nature of the stone, the development of minute micaceous flakes among the sand grains, the complex veining produced locally by infiltrated iron oxide, a thick-bedded structure exhibited by exposed bluffs, thin seams of greenish clay shale frequently magnifying the bedded appearance and the general lack of calcareous matter are characteristic of the Franconia sandstone".

And again in discussing glauconite, he (Berkey, 1898, p. 153) says that

> "an earthy, granular, bright green mineral occurs abundantly in the Dresbach formation. It is recognized as glauconite and is the same mineral that is so abundant in the St. Lawrence formation of many localities".

This leads to the conclusion that the Franconia is a white to yellow sandstone between two dominantly glauconitic formations and discourages the inclusion of other beds in the Franconia sandstone as a formation, especially since such inclusion entails the addition of other faunal zones than the *Conaspis*.

In the classification used by Twenhofel, Raasch, and Thwaites (1935) the boundary of the Trempealeau has been moved upward to include all of the upper part of the series (Table 1). The term Trempealeau had been proposed by Ulrich (1924, p. 83, 85-90) to include the Norwalk, Lodi, St. Lawrence, and local shale as he had defined them. According to the usage of Twenhofel, Raasch, and Thwaites in this classification, the Madison, Jordan, and St. Lawrence formations become mere members of the further expanded Trempealeau; the Mazomanie is dropped; the member name, Galesville, proposed by Trowbridge and Atwater (1934, p. 45) is adopted as the upper member of the Dresbach; most of the names of zones are omitted, and a number of new locality names are introduced as member names. It is clear, of course, that the several faunal zones used by these same authors in other papers lie within the members here suggested. Apparently this work formed the basis of the Conference Classification given by Trowbridge (1935, fig. 1) for use in the Guidebook of the Kansas Geological Society during their Ninth

Annual Field Conference which was conducted in the Upper Mississippi Valley.

Bridge (1937, p. 234) repeated the Conference Classification in slightly modified form. He reinstated the term Mazomanie and returned the Norwalk as a division of the Jordan sandstone. Bridge (1937, p. 233) explained very carefully that this does not imply the acceptance of the Conference Classification "either by the writer or by the United States Geological Survey." Nevertheless its publication in a Professional Paper in the same table with the United States Geological Survey recognized classifications for Missouri and for Texas places it as the most favored classification of the Upper Mississippi Valley.

THE ST. CROIXIAN IN MINNESOTA

Various member designations have often been used for the divisions of formations in the St. Croixian series of Minnesota as temporary field names but, for the most part, they have remained unpublished. The numerous changes in the St. Croixian classifications of various recent authors, or of the same author from one article to his next, suggests the changing attitude of different Cambrian students toward the problems involved. It is for this reason that the Minnesota Geological Survey has not been much interested in the abandonment of its old and fairly stable classification, particularly when that requires the dropping of formation names that have long been adjusted to the conditions as they exist in that State. A very different attitude has been taken toward the recognition of faunal zones which promise to bring about the order desired and perhaps a more usable classification for all.

The St. Croixian series of Minnesota includes dolomites, dolomitic shales, glauconites, glauconitic sandstones, shaly sandstones, and fine to coarse sandstones which here and there include conglomerates. Some of these grade laterally into beds of quite different character, and in disconnected sections the same bed may be difficult to trace or correlate. There are evidences of shore conditions and even occasional erosion surfaces. The most marked and most easily recognized of these latter is probably at the base of the Ironton member in Winona and Houston counties. Conglomerates occur at several horizons throughout the series in Minnesota, and the same is true in Wisconsin (Twenhofel, Raasch, and Thwaites, 1935, p. 1724-1727), but they appear to be of the intraformational type (Twenhofel and Thwaites, 1919, p. 628) and untrustworthy as divisional markers. Moreover the conglomerates in Minnesota are not necessarily at the same horizons as those which appear to have formed in the shallower water prevailing in Wisconsin. One such instance may be found on Stockton Hill, Winona, where two or

three different conglomerate beds occur near the base of the St. Lawrence formation as newly defined. These may or may not be present at about the same horizon in other localities, and any one of them may represent the bed that is widespread to the east of the Mississippi River at the base of the Trempealeau formation of Wisconsin (Twenhofel, Raasch, and Thwaites, 1935, p. 1690, 1705). Another instance is the conglomerate occurring in the lower part of the Tostevin quarry or within the type section of the Dresbach formation. In this case the well-formed pebbles contain the same fauna as that in the matrix itself, but this conglomerate is seldom recorded elsewhere. Numerous other instances might be cited.

The fauna of the series is particularly sensitive to the changes in sedimentation taking place from time to time, and its variation is the only reliable means of recognizing the several divisions, although sedimentary analyses have been suggestive and in many cases have been material aids. St. Croixian faunules, however, may be provokingly elusive, and this is particularly true of the portions of the series so thoroughly reworked by marine worms or other burrowing forms during sedimentation. Moreover, sandstones which are well developed at one locality may be represented by shales and glauconitic beds or even sandy dolomites at another. This is especially the case with much of the Franconia sandstone in Minnesota and the St. Lawrence formation as originally defined, although such changes are not confined to them, and some glauconite is a common constituent of the whole series. Similar changes have been traced into Wisconsin and down the river into Iowa.

PROPOSED MODIFICATIONS FOR MINNESOTA

The chief objection to the adjustment of the Minnesota St. Croixian classification to some of the changes that have been proposed by various writers and that have been adopted by the Conference Classification has been a reluctance to shift the line of contact from the top of the Taylors Falls (Goodenough) member to the top of the Bad Axe member or, as it seemed, to expand the Franconia at the expense of the St. Lawrence formation. The present members of the Minnesota Geological Survey have employed the divisions of the St. Croixian series long in use, not because they are particularly pleased with them but because they have inherited them as the official classification and no satisfactory substitute has been available. Some of the more recent changes have merit and should be adopted; others are doubtful. Minnesota geologists have preferred to continue to draw the contact between the Franconia sandstone and the St. Lawrence formation at the top of the *Conaspis* zone. Not only does this draw it at the top of the type Franconia section but also at one of the most pronounced faunal breaks in the whole series.

From the faunal standpoint they still regard this as the preferred dividing line between the two formations. Ulrich introduced a distinct name (Mazomanie) to cover most of this portion of the section claimed by two formations, and that solution had its merits although it has not met with general approval and is now usually discarded. In the interest of approximate uniformity with the usage of others in this and adjacent regions there is a feeling among the members of the committee on nomenclature for the Geological Survey of Minnesota that it may be better to discontinue the more limited use of Franconia and expand its boundary to the top of the Bad Axe member or Lower *Dikelocephalus* zone as adopted by Twenhofel, Raasch, and others. Some of the other changes suggested by these same authors or by those responsible for the Conference Classification, however, are not suited to the Minnesota development of the St. Croixian series and have not been adopted by this committee. The Jordan sandstone, for example, is one of the best aquifers in the artesian basin of Minnesota. It has a substantial thickness, is persistent with little change over the whole southeastern portion of the State, is easily mapped, and is well known to geologist, engineer, driller, and contractor alike. In fact, it is an easily recognized unit with well-marked contacts that are sometimes disconformities. Any attempt to displace a name so useful in Minnesota or to relegate it to a minor place in the classification would be unwise. Although it is true that the Jordan fauna, particularly that of the lower part or Norwalk member, bears a certain resemblance to the fauna of the Lodi member in the formation below, these two are fairly distinct as developed in Minnesota. Moreover this similarity of certain Norwalk genera to those in lower faunas goes even deeper than the Lodi, for *Dikelocephalus, Illaenurus,* and *Saukiella*—all common Norwalk genera—are said to occur in the Bad Axe member of the Franconia, and the close relationship of the Bad Axe fauna to the Trempealeau fauna has been pointed out by Twenhofel, Raasch, and Thwaites (1935, p. 1705) themselves. Probably there is no greater reason for placing the Norwalk and Lodi faunas in the same formation than there was in including the Bad Axe with the Lodi fauna—a relationship which has just been abandoned for the sake of adjustment with the Conference Classification. In the proposed Minnesota Classification the Jordan sandstone is therefore retained as an independent formation and with the same meaning as in the old classification. The name Trempealeau (Ulrich, 1924, p. 80-85), which was originally introduced as a substitute for the St. Lawrence formation, including the Norwalk, in Wisconsin, does not improve the classification for Minnesota. Moreover the Trempealeau has recently been expanded to include the entire Jordan sandstone and even the Madison (Trowbridge, 1935,

Fig. 1; Twenhofel, Raasch, and Thwaites, 1935, p. 1690, 1710-1712), if that be different from the Jordan, thus further decreasing its usefulness west of the Mississippi River. It is preferred, therefore, to retain

TABLE 2.—*St. Croixian series*

MINNESOTA			CONFERENCE (Bridge)		
Formation	Member	Fauna	Formation	Member	Fauna
	Absent ?		Trempealeau	Madison	*Plethopeltis*
Jordan	Van Oser	*Tellerina-Saukiella*		Jordan	
	Norwalk	*Eurekia-Calvinella*		Jordan (Norwalk)	*Acheilops Eurekia-Eutychaspis*
St. Lawrence	Lodi	*Dikelocephalus minnesotensis*		Lodi	*Dikelocephalus minnesotensis*
	Nicollet Creek	*Scaevogyra*		St. Lawrence	*Scaevogyra*
Franconia	Bad Axe	Lower *Dikelocephalus*	Franconia	Bad Axe	Lower *Dikelocephalus*
	Hudson	*Prosaukia-Ptychaspis*		Hudson	*Briscoia Prosaukia Ptychaspis*
	Taylors Falls	*Conaspis*		Goodenough	*Billingsella Eoorthis*
	Ironton	*Camaraspis*		Ironton	*Camaraspis*
Dresbach	Galesville	Perhaps barren	Dresbach	Galesville	
	Eau Claire	*Aphelaspis Crepicephalus*		Eau Claire	*Aphelaspis Crepicephalus Cedaria*
	Mt. Simon	Tracks and trails		Mt. Simon	

the name St. Lawrence for the remnant of that formation from which much has been ceded to the Franconia.

The classification suggested for Minnesota does not differ greatly from the Wisconsin classification as given by Twenhofel, Raasch, and Thwaites (1935, p. 1690) or the Conference Classification as given by Bridge (1937, p. 234) except that their Trempealeau is equivalent to the two topmost formations of the Minnesota classification. Table 2 gives the two in parallel columns, using a very slight modification of Bridge's interpretation of the Conference Classification.

From Table 2 it will be noted that certain member names are being used and preferred in Minnesota which differ from those of the Conference Classification. The Taylors Falls (Goodenough) member of the Franconia is typically a sandstone but it may include much shaly sandstone, glauconitic sand, and even some dolomitic beds as traced away from the type locality. It is named from the excellent outcrop of sandstone along the highway near the southwest corner of Interstate Park at the Dalles of the St. Croix River. This is the northward continuation of the type section of the Franconia sandstone and hence of the *Conaspis* zone. This member name has been used almost interchangeably with the Franconia ever since the Taylors Falls highway cut was made, and in Minnesota it is preferred to the term Goodenough member which is proposed by Twenhofel, Raasch, and Thwaites (1935, p. 1700-1701) for the same faunal zone. Goodenough is the name of a hill west of Mauston, Wisconsin, where a very good Cambrian section is exposed in a quarry and along the highway, but that portion represented by the *Conaspis* zone is now overgrown and very poorly exposed. The Taylors Falls section will remain excellent for many years.

The Hudson member consists of sandstone, sandy shales, and greensands. It carries Owen's (1852) third and fourth trilobite beds and is well developed in the sections exposed around the village of Minneiska as well as in numerous other sections both up and down the Mississippi River. Because the *Ptychaspis* fauna has been collected in that vicinity these beds have been called the Minneiska member in Minnesota and they have always been regarded as basal St. Lawrence. However, Hudson, as used by the Conference Classification, is a perfectly satisfactory name for the same beds and is adopted for Minnesota. With the change in alignment of members suggested, they become part of the middle Franconia. The Bad Axe is also a satisfactory member name. It is the upper member of the Franconia formation in the Conference Classification and is well developed in the fine section on Stockton Hill, Winona.

The Nicollet Creek member is a glauconitic dolomite which is excellently exposed along the Minnesota River at the village of Judson and across the river at Hebron where it has been quarried at half a dozen localities. It is the same as the underlying shale and the building stone or quarry beds at the old type section of the St. Lawrence formation, but the rock is more extensive and better shown at Judson and near the outlet to Nicollet Creek. This stream flows southward from Swan Lake through Nicollet township of Nicollet County and enters the Minnesota River just above the old town of Hebron or at the upper end of Judson, across the river. Sardeson (1924, p. 119) reports that he collected *Dikelocephalus* and other Lodi fossils from the dolomites and shaly

upper beds in the old St. Lawrence quarry or type section of the formation. At Judson the fauna is more limited. Below the quarry beds the shales and the conglomerates, where developed, are regarded as part of the Nicollet Creek member. In the vicinity of Judson this member is unusually well developed and may have a thickness as great as 30 to 40 feet.

The Van Oser member is a coarse massive gray to white sandstone forming the uppermost division of the series in Minnesota and is generally overlain by the Oneota dolomite. Locally, however, as along the Minnesota and Blue Earth rivers, there are thin intervening remnants correlated with other Ozarkian formations (Powell, 1935, p. 7). The Van Oser member was recognized by Winchell (1874, p. 148) and included in the Jordan sandstone. It has been regarded as the barren emergent stage of very late St. Croixian in the Upper Mississippi Valley, but its recently discovered fauna suggests the advisability of recognizing it as a distinct member of the marine Jordan, including all that portion of the formation above the Norwalk member in Minnesota. The Van Oser member is designated the *Tellerina-Saukiella* zone (Stauffer, 1939) in the Minnesota classification because of the relative abundance of these two genera.

In regard to the base of the St. Croixian series it should be pointed out that the chronologic table published in the Conference Guidebook (Raasch, 1935, Table credited to Thwaites, p. 304, column N) includes the Hinckley sandstone as the basal member of the St. Croixian series in Minnesota. The Minnesota Geological Survey map of 1932 from which the above column was taken does not include the Hinckley sandstone in the St. Croixian series. The basal member of the Dresbach formation—the Mt. Simon—may be distinguished readily from the Hinckley, and over large areas in southern Minnesota microscopic studies of the sedimentary rocks show that both the Mt. Simon and the Hinckley occur between the Eau Claire member of the Dresbach and the Red Clastic series. The Mt. Simon forms the basal member of the St. Croixian series and rests disconformably on the Hinckley from the region west of Taylors Falls, southwestward to Glencoe, and southward at least as far as Faribault and Rochester. Both the Mt. Simon and the Hinckley were penetrated by deep wells in or near St. Paul, Minneapolis, Rosemont, Glencoe, Faribault, and Rochester, and the Hinckley at least is present over the Red Clastic at Mankato. In the Bunker Hill well (Mankato) there is no sample representing the interval above the Hinckley that may be Mt. Simon, hence its presence or absence is not determined. From Faribault eastward toward the valley of the Mississippi River the Hinckley pinches out against the westward-sloping surface of the pre-Cambrian granites. In the region of Winona and Northward

the Mt. Simon overlaps the Hinckley and rests directly on the granite, as it does over large areas in Wisconsin. Along the St. Croix Valley it rests directly on the Keweenawan basalts. Atwater and Clement (1935, p. 1686) suggested that the term Hinckley be applied only to the sandstones that crop out in northeastern Minnesota. There is no structural or petrographic basis for such limitation. Both the Hinckley (Upper Hinckley of Atwater and Clement) and Mt. Simon are present in southeastern Minnesota over most of the area to which they refer. The writers believe that the exclusion of the Hinckley sandstone from the St. Croixian (Stauffer, 1925, p. 700) and its association with the Red Clastic series is correct, but the assumption that this implies the correlation of the Hinckley with the Mt. Simon (Atwater and Clement, 1935, p. 1667) is not warranted. Recent work indicates that the Hinckley sandstone is undoubtedly the equivalent of the Devils Island sandstone of Wisconsin. The beds designated Mt. Simon in Wisconsin have seldom been separated from the Dresbach formation in Minnesota. In the cases cited by Atwater and Clement, Mt. Simon was not mentioned, but the Eau Claire was doubtfully introduced to cover the lower Dresbach and therefore basal St. Croixian.

The fossils from the red shales in the Waconia well Miss Lochman has identified as belonging to the *Aphelaspis* zone, an Eau Claire faunal zone in the Dresbach formation discovered by Trowbridge near Hudson. The Waconia fossils are the only occurrence of the *Aphelaspis* fauna known in Minnesota, but it means that the Waconia well lies entirely within the St. Croixian series and that the well did not penetrate the Hinckley sandstone as originally stated (Stauffer, 1927, p. 474).

All members of the various St. Croixian formations are rather easily located in Minnesota and bear certain physical differences which are fairly well developed. Distinction, however, must be made on faunal characteristics. The Bad Axe or Lower *Dikelocephalus* zone has thus far proven unusually barren although the member is known to occur in Minnesota. As a whole the St. Croixian fauna is large. Many of the species, and perhaps some of the genera, remain undescribed. This may be true of more forms than are at present conceded to be new. Except for the Brachiopoda, preservation of the fauna is not very satisfactory, and this may be especially true of the trilobites. These are abundant but consist of the dismembered carapaces which make it difficult to determine which fragments belong together. At some places, however, fairly good material and occasionally nearly complete specimens have been collected. Wanenmacher, Twenhofel, and Raasch (1934, p. 15) have shown that intermediate faunal zones, other than those given in the table, do occur, and these are partially known in Minnesota. The faunal zones mentioned

in Table 2, however, are perhaps most important in the Minnesota portion of the Upper Mississippi Valley and for the most part have been adjusted to those used in the Conference Classification (Bridge, 1937, p. 234). A few genera named appear to be uncommon in Minnesota, but the zonal assemblage is the same, and the names proposed for such zones are satisfactory. Some of them are widespread, as suggested by Bridge (1937, p. 234) and they appear to be equivalent to certain of the faunal zones used by Resser (1938, p. 20) in his correlation table for the Southern Appalachian Cambrian formations. When the fauna is fully known it is possible that other adjustments in the classification of the St. Croixian series of Minnesota may be desirable. It is easier to find fault with the classifications now in use than it is to suggest a better one, but it is certain that too great haste in revision is undesirable.

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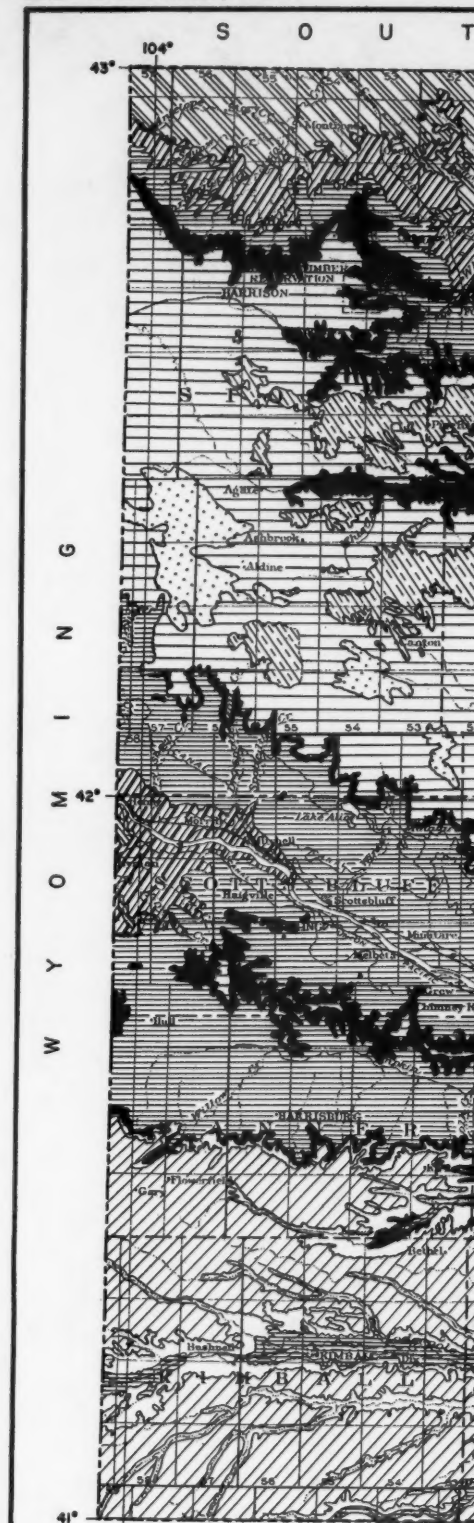
UNIVERSITY OF MINNESOTA, MINNEAPOLIS, MINN.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, DECEMBER 22, 1938.
PRESENTED BY TITLE BEFORE THE GEOLOGICAL SOCIETY, DECEMBER 30, 1938.
PUBLISHED BY PERMISSION OF THE DIRECTOR OF THE MINNESOTA GEOLOGICAL SURVEY.



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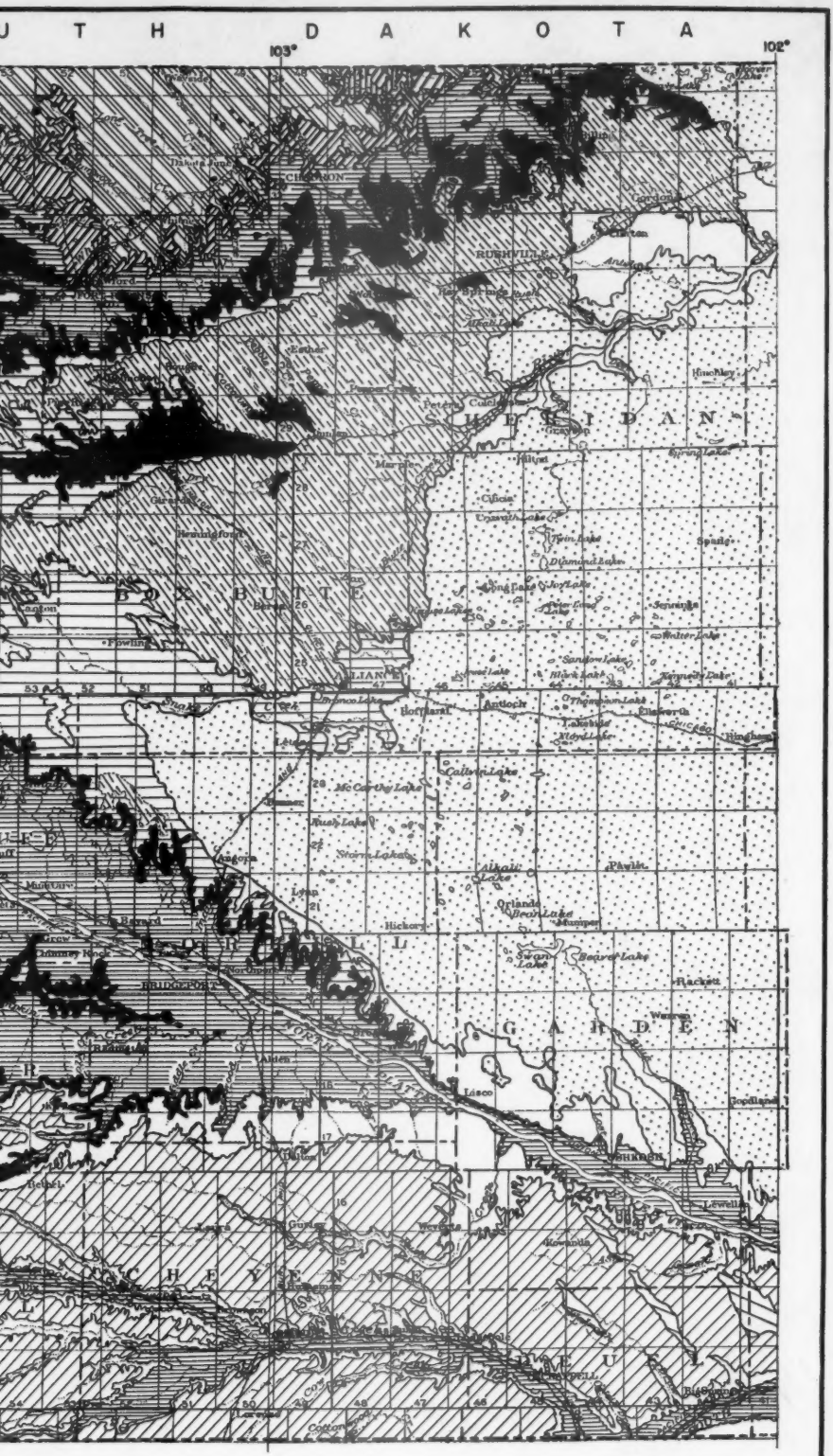

RTIARY GEOLOGIC MAP OF WESTERN NEBRASKA

e map of Nebraska by the United States Geological Survey.)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1245-1276, 1 PL. AUGUST 1, 1939

CLASSIFICATION OF THE TERTIARY SYSTEM IN NEBRASKA

BY A. L. LUGN

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1. Tertiary geologic map of western Nebraska 1245

ABSTRACT

This paper presents the results of almost continuous research by the Nebraska State Geological Survey on Tertiary stratigraphy in Nebraska during a period of more than 10 years, with special attention to the present acceptable stratigraphic nomenclature. Cogent reasons are presented for the retention of such well-known names as White River, Arikaree, and Ogallala, all now redefined and elevated to group ranking. "Loup Fork", "Loup River", "Nebraska beds", and "Republican River" are considered obsolete. The application of fossil seed zones to problems of correlation is noted. The table of Tertiary formations of Nebraska summarizes the Tertiary stratigraphy in Nebraska.

INTRODUCTION AND ACKNOWLEDGMENTS

This paper contains in the main the conclusions of the writer, who has been engaged in almost continuous research on Tertiary stratigraphy in Nebraska since 1929, regarding the classification and naming of the stratigraphic units of the Tertiary system in this State. The classification and nomenclature presented herein are now acceptable to and used by the Nebraska State Geological Survey. Stratigraphic methods can be successfully applied to problems of sedimentation sequence and faunal succession in the High Plains Tertiary deposits. The orderly arrangement and succession of "formations", so clearly set forth by Darton in several well-known publications more than 30 years ago, are realities; and it is neither necessary nor desirable to depend entirely on vertebrate fossil evidence for the determination of the age relations of either the sediments or the faunas. The multiplicity of names for "formations," "beds", "localities", and "faunas" introduced into the literature of the continental Tertiary during the past 40 years has been due to the more or less general belief that the "beds" could not be traced very far in any direction and that ordinary stratigraphic methods could not be applied to the Tertiary deposits of the High Plains. These have been erroneous assumptions, and the duplication of names is now a source of considerable annoyance.

Plate 1 is a somewhat detailed geologic map of the western part of Nebraska, in the main west of the one hundred second meridian. It has seemed advisable because of the limitations of the scale of the map to group certain Cretaceous formations under a single symbol; also, to map the Gering and the Monroe Creek formations as a unit because of the narrowness of the Gering outcrop areas. The Valentine and Ash Hollow formations are also mapped as a unit; and the Sidney gravel and the Kimball formation are shown together under another symbol. Nevertheless, the map is believed to show in considerable detail and quite accurately the areal distribution of most of the Tertiary formations.

The Brule clay extends eastward from the area included in Plate 1 along the Niobrara River Valley for more than 100 miles, and for perhaps not much more than 75 miles along the North Platte River Valley. It is present under cover throughout most of the western half of Nebraska. The Miocene formations seem to extend for only a few miles to the east in Cherry County at the north. Formations of the Ogallala group, mainly the Valentine and Ash Hollow, occur under nearly all the Sand Hills region in north-central Nebraska; and they extend far to the east of the Sand Hills under a cover of Pleistocene deposits. Outcrops of mainly the Ash Hollow and Valentine formations are fairly continuous in the

Niobrara River Valley and tributaries as far east as Knox County, and in the Republican River Valley and tributaries as far east as Red Cloud in Webster County.

Certain limitations in locating formational boundaries in the field are responsible for varying degrees of accuracy on the map. Most important are: unexposed contacts under covered slopes; similarity of lithology and lack of contrast between succeeding formations at some places, often because of the presence of reworked older material in an overlying formation; lack of exact knowledge of faunal horizons in some places; inaccessibility in some areas under existing field conditions at the time of mapping; time limitations in the field, sometimes complicated by inclement weather; in some places insufficient knowledge to date to locate all contacts with great exactness; and lastly the fact that the field mapping has been carried out by reconnaissance methods on base maps of only one inch to the mile and without the use of a plane table. The scale of the present published map is itself a limitation on the detailed accuracy of some of the formation boundaries, which had to be somewhat generalized in certain localities.

Although future more detailed mapping in local areas by more precise methods will necessitate many revisions, it is hoped that this map may be found useful. A number of small outliers of lower Ogallala (Valentine and Ash Hollow) formations occur in northeastern Box Butte County but are not shown. Certain small inliers of possible Monroe Creek formation, which may indicate anticlinal structures similar to the Agate anticline, within the Hemingford area are not all shown.

The writer has had the benefit of close cooperation with many other geologists and paleontologists during the entire period of his studies on the Tertiary system in Nebraska. Space does not permit detailed acknowledgments at this time. Dr. G. E. Condra, State Geologist of Nebraska, under whose direction this work has been done, has cooperated with the writer and given valuable counsel at all times. Dr. R. C. Moore, State Geologist of Kansas, also has cooperated and aided in many ways. Special credit for cooperation in the field work or consultation in the office or both is due M. K. Elias, C. Bertrand Schultz, Thompson M. Stout, Morris Skinner, Charles Falkenbach, F. Walker Johnson, Gilbert Lueninghoener, Emory L. Blue; Eugene Reed, Assistant State Geologist of Nebraska, and other members of the Nebraska State Geological Survey; L. K. Wenzel, O. E. Meinzer, and R. C. Cady, United States Geological Survey; E. H. Colbert, R. A. Stirton, Harold J. Cook, Ted Galusha, Jack Wilson, and indirectly to Childs Frick and E. H. Barbour; also to many local citizens in Nebraska and Kansas and to others. Much credit is due the authors of a voluminous literature dating from F. V. Hayden,

Joseph Leidy, O. C. Marsh, and E. D. Cope, to N. H. Darton, H. F. Osborn, W. D. Matthew, W. B. Scott, and many others. Wilmer R. Shirk has ably assisted in the preparation of the map.

CLASSIFICATION OF NEBRASKA TERTIARY

GENERAL STATEMENT

Classification of the stratigraphic divisions of the Tertiary of Nebraska is not simple. Many names have been applied to "beds" in numerous places within this State and in other areas. Many of these published names are recorded with only the most meager notes on the geology or even on the particular exposure in question. It is difficult in many cases to determine how the name was intended; it may have been meant to name only a fauna, or a locality, or perhaps it was applied to certain "beds".

The nonchalance with which "beds" have been given names is well illustrated by a quotation from Barbour (1917, p. 499). After pointing out that the heavy rains of spring and summer of 1915 had exposed a *Tetrabelodon* in a recent gully near Bristow, Nebraska (Boyd County), he later states,

> "The deposit is of Pliocene age equivalent to the Snake River of Cherry County, and the lower Devil's Gulch beds of Brown County. For convenience we shall call these the Bristow beds."

Some confusion is inevitable when the locality, faunal, or stratigraphic use of a name is not clearly defined by its author. The writer is primarily concerned with the stratigraphic classification of the Tertiary lithologic units in Nebraska. When the stratigraphic relationships are fully understood, the faunas can be properly and correctly related and named, whether the stratigraphic names or others are used. The Nebraska Survey, in determining its acceptance of stratigraphic names on the basis of priority, is making every effort to differentiate names which have been applied in a strictly stratigraphic sense from names which, though older, were applied in only a geographical or faunal sense and for the most part have only locality significance. It has tried to settle on a nomenclature for its Tertiary stratigraphy which will be generally acceptable and whose units are mappable. Obviously, all the faunal names cannot be applied as stratigraphic names, and, regardless of the original application of any name, only those which apply to mappable units can survive as formational or group names.

Many strictly "local" names have been applied to exposures in one sense or another, in some cases in several ways, and many if not most of them are so inappropriate or unsuitable for stratigraphic purposes that it is impossible to retain all names for stratigraphic use, and for various

reasons it is impossible to respect absolute priority in all cases. The only alternative is to try to be as considerate of the custom of priority as possible and at the same time to salvage those names which are most appropriate and suitable and which have also gained most general usage, are best understood, and have suffered the least abuse.

Also, the Nebraska State Survey considers it proper and expedient in some instances to retain a well-established name that has long been in use and is widespread in the literature in preference to some hazy, obscure name, which by some manner might be shown to have had slightly prior usage in some little-known publication, or which has not gained wide recognition and use, or has been largely abandoned. An appropriate, well-established name like Arikaree should be retained, even if it has been used in some form for some other part of the geologic section but has become obsolete or abandoned for that purpose. It is true that Cragin (1896, p. 49-52) applied the term "Arickaree shales" to some of the lower part of the Fox Hills division of the upper Cretaceous. However, Cragin himself abandoned the name, and it never gained general usage. Darton saw no objection to reviving the term "Arikaree" at a little later time and supplying his own definition (Darton, 1903, p. 17). Likewise Cragin's (1896) "Platte series" for the upper Cretaceous on a par with "Comanche series" for the lower Cretaceous never gained acceptance. This old term also has recently been resurrected, redefined, and applied to another part of the geologic section (Lugn, 1935, p. 30-31, 88-127).

The Nebraska Geological Survey adheres to a policy of cooperation in regard to all questionable problems touching matters of nomenclature, stratigraphy, and paleontology. Its policy is to reach agreement in these matters first with neighboring State surveys and then to seek approval from the United States Geological Survey. Every effort is made to define members, formations, groups, series, and systems; also, horizons and zones in the strictest sense according to the principles and rules laid down by the committee on nomenclature and stratigraphy, which was initiated by the Association of American State Geologists in 1930. This committee now represents not only the above association but also the United States Geological Survey, the Geological Society of America, and the American Association of Petroleum Geologists. A report has been published (Ashley, 1933).

PRESENT NOMENCLATURE

General statement.—A skeleton table (Table 1) of the Tertiary stratigraphic divisions now acceptable in Nebraska and Kansas and also in adjoining regions of the Great Plains is printed in this paper. Only brief

discussion can be included here. The Tertiary of this region is divisible into four natural lithologic groups, each of which represents a separate and distinct cycle of sedimentation. The White River, Arikaree, and Ogallala are familiar terms (Lugn, 1938). The White River group remains unchanged.[1] The Arikaree group has been redefined to include the Gering formation as the basal channel development at the beginning of the Arikaree cycle of sedimentation. This redefinition of the Arikaree has been announced by Schultz (1938).

White River group.—White River is the classic name for the Oligocene formations and "beds" of the Big Badlands and also for the remainder of the central and northern High Plains and in most of the Rocky Mountain region. The name has been used with little change from the early writings of Hayden (1869), Leidy, Marsh, and Cope, down to the present time (Simpson, 1933; Osborn, 1929; Osborn and Matthew, 1909). The White River group has been very conveniently and satisfactorily zoned on the basis of its contained vertebrate fossils (Osborn and Matthew, 1909; Osborn, 1929). It seems desirable to retain these faunal zones for the present at least. The White River group in Nebraska attains or exceeds a thickness of 700 feet.

The lower White River has come to be generally known as the "*Titanotherium* beds", composed of clay, silt, and channel sandstones. A conglomerate 10 feet or more thick is present at the base of the division at most places in northwestern Nebraska. This division was designated the "Chadron formation" by Darton (1899, 1903, 1905). The Chadron formation is retained with the same significance as Darton ascribed to it. It ranges in thickness from less than 50 feet to more than 100 feet and is exposed low on the slopes north of the Pine Ridge in northwestern Nebraska and in a small area in western Scotts Bluff County. The clay and silt beds grade from greenish to pink, and the sandstones may be grayish and greenish to brown.

The middle White River consists of clays, silty and sandy clays, and channel sandstones. The clays are commonly referred to as "*Oreodon*" clays, and the sandstones as "nodular sandstones" and "*Metamynodon* sandstones" (Osborn, 1929; Osborn and Matthew, 1909). The middle White River beds are as much as 125 feet or more thick where exposed in northwestern Nebraska and are predominantly pink and greenish with brownish tints in the sandstones. The upper White River division consists mainly of "*Leptauchenia* clay" containing sev-

[1] New member subdivision names for the Brule formation are now under consideration and will be introduced in forthcoming publications by T. M. Stout, C. Bertrand Schultz, and the writer.

eral layers of impure volcanic ash with the "*Protoceras* sandstones" at the base. In Nebraska this upper *Leptauchenia* zone, predominantly pink, ranges from 250 feet to perhaps 500 feet in thickness and is extensively exposed in the Pine Ridge slopes, in the North Platte River Valley, and in the Lodgepole River Valley. Darton (1899, 1903, 1905) included the middle and upper White River divisions in his Brule clay, which remains unchanged at the present. These two divisions of the Brule soon may be given formational or member designations. The Brule clay in Nebraska ranges from 350 feet to more than 600 feet in thickness.

Arikaree retained.—It seems desirable to retain and to "resurrect", if that is necessary, the well-known name Arikaree (Darton, 1899, 1903, 1905) for the main body of Miocene deposits in the "central plains area". Redefined as the "Arikaree group", it will prove to be a most useful term. It seems to be the most suitable group name for a large part of the Miocene over a very large area. Unfortunately Simpson (1933) and others (Schlaikjer, 1935) regarded "Arikaree" as an "obsolescent" term. It has never been so regarded by the writer and many other geologists in Nebraska and adjoining States, nor by the Nebraska Geological Survey. The Arikaree group now comprises the Gering, Monroe Creek, and Harrison formations.

The Arikaree "formation", as Darton (1899, 1903, 1905) called it, was defined by him as (1905, p. 176) "the principal component of the series formerly termed 'Loup Fork'" and as consisting in the main of 400 to 500 feet of fine, gray sands containing hard solid, dark gray, "tubular" concretions, occurring singly or in horizontal zones. The "pipy" concretions are present in two thick zones—in the lowermost 200 feet or more (now the lower part of the Monroe Creek formation) and in a higher zone about 100 feet thick from 100 to 200 feet below the top of the "formation" (now in the Harrison formation). Darton (1903, p. 17) stated that the *Daemonelix* of Barbour occur in the upper portion of the Arikaree. He did not at first include in the Arikaree formation anything above the *Daemonelix* beds.

The present Arikaree group, including also the Gering formation, rests with pronounced erosional unconformity on the Brule clay, mainly in contact with the *Leptauchenia* zone. The relief on the Brule surface is more than 250 feet within small areas. The deeper pre-Arikaree valleys contain the Gering formation of bedded and cross-bedded fine to coarse channel sands and gravels. The higher levels of the old Brule surface are overlapped generally by some part of the Monroe Creek formation. The maximum thickness of the Gering formation

is approximately 230 feet at a location about 6 miles south and a little west of Gering, Nebraska. Darton (1905, p. 177) stated that,

> "it is believed that the Gering beds represent the first deposits of Arikaree times, separable only in areas where they consist of coarse material laid down in channels,"

but he did not include it with the Arikaree. The Gering channel deposits are best known and perhaps most fully developed in the North Platte River Valley in Scotts Bluff and Morrill counties and in the slopes of the Pine Ridge across northern Sioux and Dawes counties. The Gering horizon seems to change to a floodplain facies closely resembling the Monroe Creek formation from Chadron eastward across northern Sheridan County.

The Monroe Creek formation was defined by Hatcher (1902) as a lower subdivision of the Arikaree from outcrops in Monroe Creek canyon in the Pine Ridge a few miles north of Harrison, Nebraska. The Monroe Creek formation along the Pine Ridge from Harrison to the Crawford vicinity rests conformably on from 100 to 200 feet of Gering channel deposits. The thickness of the Monroe Creek formation in this area ranges from about 285 feet to more than 360 feet. The lower part, from 185 to about 220 feet, is composed of fine- to medium-textured gray sand, massively bedded for the most part, and permeated by large, heavy, "pipy" concretions, the "lower" pipy zone. Some cross-bedding may be observed in the lower part of the formation.

The upper part of the Monroe Creek formation, 100 to nearly 150 feet thick, consists of compact fine sandy silt and clay, pinkish to buff, containing many layers of more or less cemented concretions and many isolated nodular concretions. The more solid concretion layers, about a foot thick, are quite flat and horizontal along their upper surfaces; but peculiar pendant irregular concretionary structures extend downward for a foot or more from the lower side. Many of the isolated nodular concretions are elongated and "characteristically upright in position" (Cook, 1915). The concretions in the upper part of the Monroe Creek are characteristically different from the heavy, "pipy" ones in the lower part of the formation or the "pipy" concretions in the lower part of the next overlying Harrison formation.

The Harrison formation first was differentiated as an upper subdivision of Darton's Arikaree by Hatcher (1902). It rests conformably on the Monroe Creek formation, and the top of it is at the general level of the High Plains in the vicinity of Harrison, Nebraska. It consists of about 200 feet or slightly less of fine, mostly unconsolidated, for the most part massive, gray sand. The lower part, about 100 feet thick, also is permeated by horizontal "pipy" concretions, the "upper" pipy zone, similar to but not as large and "heavy" as the big "pipy"

structures in the lower part of the Monroe Creek formation. The upper 100 feet of the Harrison formation contains the *Daemonelix* structures and considerable "fossilized" vegetal remains, which take a stringy, rod-like, or nodular form. Less of this stringy material is found in the lower part associated with the large pipy concretions. A thin characteristic agatized limestone bed caps the formation. Hatcher's Harrison did not include the so-called "upper Harrison" (now Marsland formation), which came into use a little later (Peterson, 1906; Cook, 1915; Schultz, 1938).

Thick fills of channel sediment, 50 to more than 150 feet thick, occur in the Harrison formation and in general resemble quite closely the lithology of the channel deposits of the older Gering and those of younger formations, as in the Sheep Creek, Valentine, Ash Hollow, or Snake Creek. The well-known fossil bone deposit at Agate is a part of a Harrison channel fill near the middle of the formation. Another deep channel fill, perhaps stratigraphically a little higher than that at Agate, occurs closely associated with fossil beds a few miles south of Angora; and a thick Harrison(?) fill seems to extend across the southern part of Box Butte County.

Hemingford group and Marsland formation.—The Hemingford group is a new division suggested by C. Bertrand Schultz, Assistant Director, Nebraska State Museum, and concurred in by the Nebraska Geological Survey and the writer. It includes the Marsland and Sheep Creek formations. The Marsland formation is a recent term given by Schultz (1938), who has stated fully the basis for the use of this term and the redefinition of the Arikaree. He writes:

> "To the deposits ('upper Harrison beds') which immediately overlie the Arikaree group and which are faunally and lithologically distinct from the typical Arikaree, the writer suggests the name Marsland formation. This formation is best exposed in Nebraska in the region about Marsland (southwestern Dawes County) along the Niobrara River where it includes some 150 feet of buff and gray, soft sandstones. The Marsland consists, in part, of valley fills, and in places seems to mantle the slopes of certain large valleys. The upper part of the Marsland formation in this region is more gritty and is mostly buff-colored. The fauna of the Marsland formation seems to be intermediate between that of the Harrison formation and the overlying Sheep Creek formation and perhaps should provisionally be considered as the lower part of the upper Miocene. The characteristic fossils are *Merycochoerus, Merychyus, Aletomeryx, Oxydactylus,* and advanced *Parahippus,* or (and) *Merychippus,* etc."

The Harrison formation is to be understood in the sense in which it was originally defined by Hatcher (1902). The Marsland formation, the buff to reddish-brown beds which overlie the Arikaree group, was never included in the Harrison formation of Hatcher; neither was the Marsland a part of Scott's "Nebraska beds", as Hatcher thought, nor was it included

in Darton's Arikaree (Darton, 1903) except as he accepted Hatcher's subdivisions in a later report (Darton, 1905). Darton states:

> "According to Hatcher (1902) the Arikaree formation in the western portion of Pine Ridge is divisible into three distinct horizons, comprising the Monroe Creek beds, which overlie the Gering sandstones north of Harrison, Nebraska; the Harrison beds, which constitute the surface of Pine Ridge in the vicinity of Harrison; and the Nebraska beds of Scott, which occur along Niobrara River south of Harrison."

It should be evident that the stratigraphic relationships of the Marsland beds have not been very clearly understood. Hatcher may have meant the beds now called the Marsland formation, but they were not the "Nebraska beds" of Scott. The term "Nebraska beds" (perhaps correlative of the Valentine formation) is no longer in use (Simpson, 1933).

However, even though Darton (1905) thought that Hatcher (1902) meant to include the "Nebraska beds of Scott, which occur along the Niobrara River south of Harrison", in the Arikaree, reference to Hatcher's (1902) original paper does not seem to substantiate this conclusion. He states clearly of the Arikaree that, "it is lithologically and faunally divisible into two easily distinguishable horizons". He undoubtedly meant to include in the Arikaree the Monroe Creek and the Harrison beds (in the Hatcher sense) but he apparently did not intend to include the "Nebraska beds" of Scott, as Darton seems to have thought. Hatcher suggested in the same paper that the Arikaree should include only the Gering and Monroe Creek beds; and that the Harrison should be placed with the Nebraska beds (as he then believed them to occur) and the Ogallala to constitute a redefined Loup Fork group. Hatcher's suggestions apparently were not accepted and did not come into general use.

Therefore, these buff to reddish-brown beds have never had a proper name until the term Marsland was proposed recently (Schultz, 1938). The application of the name "upper Harrison" to these beds by Peterson (1906) and others (Cook, 1915) in more recent years has been unfortunate because these beds (Marsland) have no very close stratigraphic or faunal relation to the true Harrison (in the Hatcher sense) and are separated from the underlying Harrison formation by the most significant and important structural and erosional unconformity within the Miocene series in western Nebraska. The time of the major folding which produced the Agate anticline and certain other structures in western Nebraska seems to have been post-Harrison pre-Marsland.

Snake Creek and Sheep Creek formations.—These formations were defined and named first by Matthew and Cook (1909) and further described and discussed in several papers (Cook, 1915; Matthew and Cook, 1918; Matthew, 1923, 1924). The type locality for both of these formations

is in south-central Sioux County, Nebraska. The Sheep Creek beds were said to lie unconformably on

> "Lower Miocene, equivalent to the *Daemonelix* Beds of the Niobrara valley" [Harrison formation]. . . . They consist of soft fine-grained sandy 'clays' of a light buff color, free from pebbles, and containing harder calcareous layers"

(Matthew and Cook, 1909, p. 362).

Matthew and Cook (1909) considered the Snake Creek beds "the remains of a formation" which they regarded as an "outlier of the Ogallala". The Snake Creek formation was said to lie on "the eroded surface of the Sheep Creek beds" and it comprised mainly channel-fill deposits, lacking in the usual amount of calcareous cementing material typical of the "mortar beds" farther south. Of the Snake Creek beds,

> "they are composed of a clean sand, with gravel scattered through it, especially in certain layers, and mantling the eroded surface of the Sheep Creek beds"

(Matthew and Cook, 1909, p. 363).

It is unfortunate that the apparent clarity with which Matthew and Cook (1909) at first differentiated these two formations did not always characterize later writings (Matthew, 1923, 1924). Confusion has resulted in the minds of other workers (Simpson, 1933), especially because of the "mixed" faunal lists which have resulted from mixed collections from the Sheep Creek—Snake Creek locality.

It seems evident that the typical "Snake Creek" formation, or at least the so-called "upper Snake Creek", is definitely correlative with some part of the upper Ogallala group. Unfortunately, "Snake Creek" has been incorrectly (?) applied to beds, mainly channel fills, as late as Pleistocene and to other beds as old as Miocene. Much of the Miocene "Snake Creek" or perhaps the so-called "lower Snake Creek" seems to be in part not Snake Creek at all but Sheep Creek channel beds in place and in proper stratigraphic sequence. Also, some of the Miocene and "Sheep Creek" vertebrate fossils collected from "Snake Creek" channel deposits have come from very large blocks of Sheep Creek formation, which were broken away from the banks of the Snake Creek streams. These large blocks of rock have dimensions of 10 to 20 feet or more and were never broken up or disintegrated by the Snake Creek rivers. They lie buried in the younger silt and sand, the true Snake Creek sediment, which is always channel-fill material.

The Sheep Creek formation consists of channel gravels and sands, fine silty sand, silt and clay, and harder caliche beds of widely ranging textural characteristics, all of which fill for the most part narrow valleys or ravines to depths ranging from 30 or 40 feet to 140 feet or more. These deposits, many of which are isolated and unconnected, are widespread throughout quite a large area in northwestern Nebraska; and, in one area

in the central part of the south half of Sioux County, the Sheep Creek formation extends continuously as a basin deposit of channel, floodplain, and slack-water sediments over about 25 square miles. This seems to be the most extensive and most continuous known occurrence of the Sheep Creek. The areal extent of the Sheep Creek formation no doubt has been larger than at present. The discontinuity of the formation throughout the area of its present distribution probably is due in part to erosion subsequent to deposition.

The total thickness of the Sheep Creek formation exposed above the unconformable contact on the Harrison formation in the type locality, *Pliohippus* and *Aphelops* draws in southern Sioux County, Nebraska (Matthew, 1924a), is more than 200 feet. Correlative and stratigraphically overlapping exposures in the southeastern corner of Dawes County, along Pepper Creek and Sand Canyon, clearly indicate that the Sheep Creek stratigraphic and sedimentation unit—the fundamental formational entity—includes about 70 feet of additional beds which are stratigraphically higher than the uppermost Sheep Creek exposures in the type locality. This increases the stratigraphic thickness of the Sheep Creek formation to at least 270 feet, not including a still higher "member" to be described. It has been mainly the upper part of the Sheep Creek (mostly from about 100 feet below and on a distinctive dark-gray volcanic ash 2 to 3 feet thick) which has been called by some the "lower Snake Creek beds" (Matthew, 1924a) but which in reality constitutes the latest deposits of Sheep Creek sedimentation.

The Sheep Creek formation as now understood can be zoned on the basis of three or four fairly distinctive species of fossil grass seeds, genus *Stipidium,* with at least one species of fossil *Berriochloa,* into at least three floral or fossil seed zones. No physical field evidence or faunal data are known which would justify the subdivision of the Sheep Creek formation into new and smaller formational divisions. Floral and faunal zones and member subdivisions may be justifiable. Complete details on the zoning of the Sheep Creek formation will be published in a forthcoming bulletin of the Nebraska State Geological Survey.

The Sheep Creek formation, according to much of the literature, would seem to be of medial to late Miocene age. That it is of late upper Miocene age seems quite probable at present. Many vertebrate fossils supposedly collected in Sheep Creek beds have actually come from the underlying older Marsland formation or from the still older Harrison. This has contributed to the difficulty of dating the true Sheep Creek deposits. Confusion as to the age and possible contemporaneity of parts of the Sheep Creek formation with channel fills sometimes regarded as "Snake Creek channels" has been due to a mistaken notion that all the "channel"

deposits in the type area (Matthew, 1923, 1924) were Snake Creek and that the slack-water and floodplain deposits were Sheep Creek; that is, the so-called "Snake Creek" (channel-filling deposits) and the "Sheep Creek" (slack-water and floodplain materials) were only facies developments of the same sedimentary deposit, a conception entirely contrary to the first definition and description (Matthew and Cook, 1909). Matthew (1923), writing about "The Snake Creek fossil quarries", states:

> "The pockets from which the great bulk of the material has come are channel-beds excavated in and partly contemporary with fine-grained muddy sandstones, to which the name of Sheep Creek beds was applied when first found. These appear to be back-water sediment corresponding in age to horizons 1 and 2 of the channel-bed series [called by Matthew and Cook "Snake Creek" formation]. No back-water facies has been recognized for the Pliocene channel-beds (No. 3)" [the true Snake Creek channels at the top of the deposits].

Later Matthew (1924), writing about the relations of the Snake Creek and Sheep Creek beds, states that,

> "the two [Snake Creek and Sheep Creek formations] representing different facies of the same formation or sequence of strata, in part contemporaneous, rather than two distinct formations. The Snake Creek beds are channel-fillings throughout (except near the top), and they fill and overlie eroded channels in the Sheep Creek beds, which are rather a fine-grained uniform back-water or flood-plain formation. But the stratigraphic and faunal evidence jointly show that the flood-plain beds are contemporary with at least the two older horizons (Miocene) of the channel beds. The third faunal horizon of the channel-beds [Snake Creek in original sense—1909] has no recognized flood-plain correlative."

Such an intermingled complex of contemporary sedimentation facies, as noted above from Matthew's papers, cannot be recognized as two separate and distinct stratigraphic formations by any criteria known to the writer. However, the original definition of the Snake Creek and Sheep Creek formations (Matthew and Cook, 1909) may be essentially valid, and it is in this sense that the writer proposes to continue the use of these names to apply to correctly differentiated formations. With this in mind, it should be noted that most if not all the so-called "Snake Creek" faunas from horizons 1 and 2 were collected from channel fills within the Sheep Creek formation. The Sheep Creek formation consists of channel fills, slack-water, and floodplain sediments (includes "Snake Creek" faunal horizons 1 and 2); whereas the higher and considerably later (middle to late Ogallala) true Snake Creek formation in the 1909 sense are all channel or ravine fills. These Snake Creek channel deposits may have filled and choked the Pliocene streams, which may have supplied the materials for quite widespread gravel and sand beds in the Ash Hollow or even the Sidney formation in the Ogallala group throughout areas to the south and southeast.

After the Sheep Creek "channels" or narrow valleys and ravines which had been eroded into older formations—in all known cases either the Harrison or the Marsland—had been filled with sediment, quite ex-

tensively distributed beds of floodplain "marl" and eolian silt were deposited. The distribution of this heretofore unrecognized member of the Sheep Creek formation seems to have been nearly the same as that of the Marsland formation, and at present it overlies and extends beyond the limits of most of the Sheep Creek valley fills more or less throughout the areal extent of the Hemingford group, and in a few places it even rests on the Harrison formation.

The lower part of this upper clay member of the Sheep Creek formation consists of reddish-brown clay, quite plastic, with much colloidal material, in all about 30 to 35 feet thick. A middle zone (10 to 12 feet) of this lower clay is somewhat more silty and gritty, with silico-calcareous concretions throughout which greatly resemble loess concretions. This middle silty zone is somewhat lighter, being more pinkish to slightly greenish. Next above the lower reddish-brown clay occurs a middle zone of silty grayish and brownish lumpy to blocky fine-grained sandstone which includes a bed of bluish-gray volcanic ash at some places. This middle zone is fossiliferous (mostly fragmental bones) in places. This middle sandstone zone also is about 35 feet thick. A third and higher zone of greenish to reddish-brown silty, loess-like clay, similar to the lower clay zone, only about 15 feet thick occurs at a few places at the very top of this new "upper" member of the Sheep Creek. The total maximum thickness of this new member of the Sheep Creek formation is about 85 feet, and it is most fully developed in T. 29 and 30 N., R. 47 W., in southeastern Dawes County, Nebraska, a few miles northeast of Dunlap.

This upper member of the Sheep Creek generally has been included with and measured as a part of the Marsland formation, on which it rests unconformably at most places, where no separating Sheep Creek valley fill deposits occur to reveal the true relationship. No stratigraphic name is proposed for this new member by the writer. The privilege and responsibility of supplying a name for this formational subdivision belong to Richard C. Cady, of the United States Geological Survey, who is its discoverer and who is senior author of a publication in preparation on the geology and ground-water resources of Box Butte County, Nebraska.[2]

Ogallala redefined.—The Ogallala, originally named and defined by Darton (1898, 1903, 1905), is redefined as a group of four definite and mappable formations. The formational names—Valentine, Ash Hollow, Sidney, and Kimball—were not the choice of the writer alone, nor of any one individual. These names were agreed upon first in conference early in March 1936 in the State Geologist's office at Lincoln, Nebraska, by

[2] A cooperative project of the United States Geological Survey and the Conservation and Survey Division of the University of Nebraska.

G. E. Condra, State Geologist of Nebraska; R. C. Moore, State Geologist of Kansas; Maxim K. Elias, of the Kansas State Geological Survey, and the writer. Agreement on Ogallala nomenclature also had been reached with C. Bertrand Schultz, Assistant Director, Nebraska State Museum.

The appropriateness of the name Ogallala for the very widely distributed group of formations now included in this division seems not to have been fully appreciated. Ogallala has been applied more inclusively to all the Pliocene deposits of the Great Plains than any other name, which has gained recognition for any certain part of these deposits, except perhaps the very general and obsolete "Loup Fork". Ogallala seems to be the only name that actually has been applied to Pliocene formations from the oldest (Valentine) to the highest and youngest beds, the "Algal" limestone of Elias (1931), from South Dakota to Texas. Other terms, such as Valentine, Republican River, and Devil's Gulch include only a part of the group or formation, as the Ogallala has formerly been called, and all these named "beds" are for the most part restricted to the lower half or less of the "Ogallala". Not one of these other names could be redefined to include the entire group of formations. As a matter of fact, most of our knowledge of the Pliocene series in Nebraska and adjacent areas has been only piecemeal and mostly limited to the lower beds. Very little has been known of the character and faunal content of the upper part of the Ash Hollow, the Sidney, or the Kimball formation. Most of the faunal collections from the higher formations or beds have never been correctly "tied in" to the general stratigraphic section of the Ogallala group.

The oldest and lowest formational subdivision of the Ogallala group is the Valentine formation, which was introduced by Barbour and Cook (1917). The validity and appropriateness of this name have been well stated by Johnson (1936). The writer is in full agreement with others (Johnson, 1936; Simpson, 1933) that the name "Valentine" is correctly applied only when used for the older unconsolidated sands and gravels in the basal part of the Ogallala.

A redefinition of "Valentine" has been proposed recently, thereby introducing a "Valentine Problem" into the literature. This incorrect application of the name Valentine in stratigraphic and paleontologic literature seems to have appeared first in papers by Stirton (1933) and Stirton and DeChardin (1934). Shortly thereafter Stirton and McGrew (1935) definitely proposed a redefinition of Valentine. Johnson (1936) published a lucid and complete explanation of the problem. His paper should have settled the matter for all time, but a little later Stirton (1936) published another paper in which he continued the misapplica-

tion of Valentine. The problem has been revived again by McGrew and Meade (1938) who persist in applying "Valentine" to a part of the geologic section which was not included in the original "Valentine beds" of Barbour and Cook (1917). The general disagreement with the proposal to redefine the Valentine is clearly indicated in papers by Lewis (1938), Colbert (1938), Johnson (1938), and the writer (Lugn, 1938).

Nevertheless, Valentine is a very satisfactory name for all the more or less unconsolidated, fine gray sands, 175 to 225 feet thick, which constitute the lowest and oldest part of the Ogallala group. The Valentine formation occurs under the typical Ogallala "mortar beds." The "cap rock bed", a field term mentioned by Johnson (1936), is the lowest part of the "mortar beds" division of the Ogallala group. It is also almost exactly the equivalent of the *Krynitzkia* fossil seed zone of Elias (1932, 1935; Chaney and Elias, 1936). It contains the "upper" of Stirton and McGrew's (1935) three faunas, for which they proposed the name "Valentine" in a redefined sense. The Burge channel member of the Valentine formation occurs at the very top of the Valentine below the *Krynitzkia* seed zone, and the "Valentine fauna" comes from lower down in the formation (at about 80 feet above the Valentine-Brule clay contact southeast of Valentine, Nebraska).

The "mortar beds" division of the Ogallala group, which in the main is the only part exposed[3] in the "type locality" (?) near Ogallala, Nebraska, as defined by Elias (Stirton, 1936; Hesse, 1935), is known as the Ash Hollow formation, 100 to 250 feet thick. It contains the "cap rock bed" of the *Krynitzkia coronoformis* fossil seed zone, the "fragmental" vertebrate fossil zone which is the third and highest faunal zone of Stirton and McGrew (1935)—mistakenly called by them "Valentine",—in the lower part of the formation. The remainder of the Ash Hollow formation includes most of the *Biorbia fossilia* fossil seed zone of Elias (1932, 1935; Chaney and Elias, 1936). This part of the formation also contains other faunal zones; and in the Valentine, Nebraska, area, near Burge, farther east in northeastern Cherry County, northern Brown County, and in Keya Paha County, the lower part of the *Biorbia* zone contains an abundant vertebrate fauna, which is higher than any mentioned by Stirton and McGrew (1935). The association of *Biorbia fossilia* with this higher fauna was discovered by Morris Skinner and the writer in 1935. Faunas correlative to the two in the lower part of the Ash Hollow formation in the Valentine, Nebraska, area and other still higher faunas are known from other areas in southern and southwestern Nebraska.

[3] The unconsolidated Valentine sands are now known to occur here under the "mortar beds," from data secured by test drilling.

The type section of this formation occurs in exposures in Ash Hollow Canyon southeast of Lewellen, Nebraska. The beds belonging to this formation in Ash Hollow Canyon, above about 55 feet of Brule formation, consist of layers of gravel and sand, silt, and fine sandy clay, with some beds of volcanic ash, all more or less indurated into hard caliche beds at fairly regular intervals. The total thickness of the beds believed to belong to the Ash Hollow formation, from the top of the Brule clay to the base of the overlying Sidney formation, is from 250 to 265 feet. This thickness includes a layer of conglomeratic limy sand and gravel which appears to lie below the *Krynitzkia* fossil seed zone, and this bed may be found to correlate with the Burge channel member of the Valentine formation. The Sidney gravel formation and remnants of the Kimball formation (Lugn, 1938) occur at elevations above the top of the Ash Hollow under the High Plains tableland level to the south of the canyon.

It has come to the writer's attention recently that the "Ash Hollow formation" is not a new or even recent name, as was at first supposed (Lugn, 1938). The term Ash Hollow formation was applied to exactly the same beds in Ash Hollow Canyon, the same location and exposures noted above, and they were said to "attain a thickness of over 250 feet" first by Henry Engelmann in 1858 or 1859 (Engelmann, 1876). It is a source of gratification that the recent redefinition (or reapplication) of the Ash Hollow formation by the writer and others matches perfectly in every way—lithology, thickness, location, and type section—the original definition of the "Ash Hollow formation" by Engelmann.

The third formation of the Ogallala group is known as the Sidney gravel from occurrences at Sidney, Nebraska. It ranges in thickness from 15 to 50 feet and is widespread in southwestern Nebraska, northeastern Colorado, and in parts of western Kansas. It is the upper part of the "*Biorbia* fossil seed zone", although *Biorbia* also occurs in the Kimball formation.

The type exposure of the Sidney gravel is located in the high bluff at the north side of the town of Sidney, Nebraska, a few rods west of the elevated water tanks which are a part of the Sidney water supply system. The gravel bed at this location is 20 feet thick and consists of crystalline sand and gravel ranging in texture from fine and medium river sand to pebbles and cobbles as much as 4 to 6 inches in diameter. Many of the pebbles are covered with a thin dark coating of mineral and clay which give to the deposit a characteristic darkish appearance in many fresh exposures. The gravel bed rests on a hard caliche sandstone or "grit" layer, belonging in the Ash Hollow, which is commonly quite characteris-

tically pitted with old "fossil" potholes, some of which are as much as 5 or 6 feet deep and several feet in diameter. Many potholes may be observed below the gravel at the type exposure, but none at this place seem to attain the maximum size observed at a few other exposures.

The Kimball formation occurs typically developed above the Sidney gravel under the High Plains level for many miles north of Sidney. However, the relationships of the uppermost Ash Hollow beds, the Sidney gravel, and the beds of the Kimball formation, in a relatively small area north and northeast of the "water tanks" location at Sidney, are complicated by a zone of normal faulting which trends about west-northwest, more or less parallel to a conspicuous drainage.

The Sidney gravel in the main is a quite widespread sheet-like complex of channel deposits, but there are small areas where it is not developed; and, in such places, the Kimball formation rests directly on the Ash Hollow formation commonly with apparent lithologic continuity. The Sidney gravel is "pinched" out in this manner in one small area a few miles west of Sidney, Nebraska. Other less important and less widespread channel gravel deposits, similar in lithology to the Sidney gravel, also occur at lower horizons in the Ogallala group, especially in the middle and upper parts of the Ash Hollow formation.

The Sidney gravel is not fully differentiated from the Pleistocene "high terrace" in southern Garden, Deuel, and perhaps in parts of Keith counties. In this area, mainly between the North Platte and South Platte rivers, the gravel-covered Pleistocene high terrace level accords topographically almost exactly with the stratigraphic position of the Sidney gravel in the Ogallala group. Some of the Sidney gravel has been reworked on this Pleistocene level; and, since the gravels of both ages are unconsolidated, it is impossible to make precise differentiation. At some locations, Pliocene (Sidney) fossil bones have been found in the lower part of the gravel, indicating that at these places at least the lower part of the gravel is Sidney in place. Also, Sidney gravel occurs undisturbed in small areas still capped with Kimball caliche and algal limestone zones. Pleistocene fossil bones have been found in the upper levels of the gravels at some points where not overlain by Kimball, and this indicates the position of the Pleistocene high terrace level and the reality of the reworking of part of the Sidney gravel.

A fourth and uppermost formation of the Ogallala group is known as the Kimball formation from its typical occurrence at the highest remnant levels of the High Plains in Kimball County, Nebraska. The thickness of the Kimball formation ranges from 25 to 50 feet where present in its full development and consists of silt, clay, and fine sand, partly cemented

with caliche, with one or two algal limestone beds at the top. It is pinkish to reddish and contains the fossil seeds of the genera *Echinochloa, Panicum,* and *Biorbia* (Elias, 1932; Chaney and Elias, 1936). The most striking feature of this formation is the occurrence, at the top of an algal limestone *(Chorellopsis bradleyi* Elias) (Elias, 1931, p. 136-141). The formation is also widespread and generally is not overlain by later deposits, except small amounts of wind-blown silt or loess material.

The Kimball formation in general consists of three beds or zones. The lowest bed, resting with apparent conformity on the Sidney gravel at most places, is a grayish and pinkish "grit" or caliche sandstone, composed of fine to coarse sand, with small granule pebbles in some places. It may contain nodular bands of milky chert or chalcedony, generally fairly hard and indurated, forming a more or less prominent ledge. This lowest zone ranges from 5 or 10 to 15 feet in thickness. The middle bed of the Kimball formation, 18 to 22 feet thick, consists of fine silty sand, pinkish and brownish, and for the most part soft. The uppermost or algal limestone zone generally is one conspicuous bed of whitish hard algal limestone 2 to 3 feet thick; but it may consist of two or even three algal limestone layers with interbedded soft pinkish to grayish fine silty sand and marly clay, in all ranging up to 12 to 15 feet in thickness. Milky chert also may be a constituent of this uppermost bed of the Kimball formation, and minute siliceous structures strongly suggestive of spore cases of the algal genus *Chara* are also present.

The lower part at least of the Kimball formation generally contains *Biorbia* fossil seeds and certain other fossil seeds, as noted above, which Elias believes to be restricted to the uppermost beds of the Ogallala group. A typical exposure of the Kimball formation occurs at the High Plains level about 2 miles south of Kimball, Nebraska, in the vicinity of the adjoining corners of sections 5, 6, 7, and 8, T. 14 N., R. 55 W.; and many other good exposures occur from this point southward into Colorado. Typical development of the Kimball formation, where the beds are well exposed, may be seen also at these locations: south of Harrisburg, Nebraska, in Banner County, in the S. E. ¼, sec. 26, T. 18 N., R. 56 W.; at 9 miles east of Sidney, Nebraska, in Cheyenne County, the S. W. ¼, sec. 23, T. 14 N., R. 48 W.; near the Lone Star School in Colorado, 20 miles south of Kimball, Nebraska; and at innumerable other locations in southwestern Nebraska, northeastern Colorado, and western Kansas.

The stratigraphic table of Tertiary formations of Nebraska (Table 1) provides a brief summary of the Tertiary stratigraphy in this State and a résumé of the classification and nomenclature of these formations, in use at the present time.

TABLE 1.—*Tertiary formations of Nebraska* *

PLIOCENE

1. Ogallala group 300′-500′
   (1) Kimball formation 25′-50′
   Echinochloa, Panicum, and *Biorbia* fossil seeds, and Algal limestone.
   (2) Sidney gravel 15′-50′
   Upper part of *Biorbia* seed zone.
   (3) Ash Hollow formation 100′-250′
   Main part of *Biorbia* fossil seed zone, and several faunal horizons in *Biorbia* zone; *Krynitzkia* seed zone in the lower part ("cap rock bed"), "fragmental" vertebrate fossil zone ("Valentine" of Stirton and McGrew).
   (4) Valentine formation 175′-225′
   Stipidium fossil seed zone; Burge channel member and the "Burge" fauna at the top; the "original" Valentine fauna occurs well down in the formation.

Snake Creek channels 0′-75′

(?) UNCONFORMITY

MIOCENE

2. Hemingford group 250′-400′
   (1) Sheep Creek formation 140′-300′
   Stipidium and *Berriochloa* fossil seeds.
   Unconformity
   (2) Marsland formation (old "Upper Harrison") 125′-200′

UNCONFORMITY

Note: Horizon of the folding of the Agate Anticline and many other structures.

3. Arikaree group 700′-800′
   (1) Harrison formation 200′
   Daemonelix beds at the top.
   (2) Monroe Creek formation 375′
   (3) Gering formation 100′-200′

UNCONFORMITY

OLIGOCENE

4. White River group 500′-700′
   (1) Brule clay 500′-600′
   Leptauchenia and *Oreodon* beds with zones of channel sandstones.
   (2) Chadron formation 50′-150′
   Titanotherium beds.

UNCONFORMITY

Mainly Cretaceous shale formations: Pierre, etc.

NOTE: *Celtis* seeds occur throughout all of the Tertiary formations noted above. *Celtis hatcheri* has been described as from the White River group; most of the *Celtis* stones from higher formations are generally referred to the species *willistoni.* The genus *Stipidium,* perhaps several species, occurs also in the higher formations of the Ogallala group associated with the "index" fossil seeds of these zones. The Miocene-Pliocene contact is still undecided upon as indicated above.

* Reproduced with slight alterations from the writer's paper (Lugn, 1938) in the American Journal of Science.

AGE OF TERTIARY FORMATIONS IN NEBRASKA

It is not the intention of the writer in the present paper to try to fix definitely the age of the Tertiary formations in Nebraska with respect to the accepted epochs of the Tertiary period, except in a general way. At present it is impossible to assign exact age to any of the formations

discussed herein, until the vertebrate paleontologists can agree on criteria to be applied in determining the limits of all the Tertiary epochs.

It has been possible to determine the age relations of the several groups and formations largely by field study, and the sedimentation sequence and stratigraphic succession of beds have been worked out with considerable certainty. The areal distribution of the stratigraphic groups and formations has been mapped in some detail. The relative ages and positions in the geologic column of all the named sedimentary and stratigraphic units are now understood with greater accuracy than ever before. These are the results in part presented in this paper, and they are considered of far greater importance at present than the assignment of exact age to any particular beds, desirable as that also may be. But, until agreement is reached on what criteria shall be used to differentiate, for example, Miocene from Pliocene, age assignments to formations cannot be made with any certainty of agreement among either stratigraphers or paleontologists.

No deposits of Eocene age are recognized in Nebraska. Considerable confusion exists in regard to the age of the Marsland and Sheep Creek formations. This has been discussed earlier in this paper, but the confusion regarding the age of the Sheep Creek formation is due largely to the fact that many vertebrate fossils supposedly collected from the Sheep Creek beds actually have come from the older, underlying Marsland formation, or even from the still older Harrison beds. The causes of confusion regarding the age of the Snake Creek beds have been noted, but fundamentally the main source of difficulty has been the fact that the Snake Creek faunal horizons 1 and 2 actually occur in the Sheep Creek formation and only faunal horizon 3 is in the Snake Creek formation. The uncertain age of the Valentine formation is due largely to lack of agreement on the criteria to be applied in differentiating Pliocene from Miocene.

It will be necessary to know more about the sedimentary succession and stratigraphic relations of all the beds throughout the Tertiary system than is understood at present, in order to understand the age relations and succession of fossils and faunas, before agreement can be expected on the exact age and epoch divisions of the Tertiary period. Therefore, only very tentative age assignments are made for the Tertiary formations in Nebraska in this paper. Table 2 represents only the *tentative opinion at the present time* of the Nebraska State Geological Survey, the Nebraska State Museum, the writer, and numerous other geologists and paleontologists. The table should not be regarded as indicating any final or settled opinion. It is subject to revision as new data may require at any time.

THE "HARRISON" IN THE GOSHEN HOLE AREA

The geology of an area lying mostly west and northwest of Scotts Bluff County, Nebraska, known as the "Goshen Hole Area, Wyoming",

TABLE 2.—*Tentative age assignments to Tertiary formations in Nebraska*

PLIOCENE	Ogallala	Kimball formation (very late upper Pliocene) Sidney gravel (upper Pliocene) Snake Creek formation (middle to late Pliocene) Ash Hollow formation (middle Pliocene) Valentine formation (early Pliocene)
MIOCENE	Hemingford	Sheep Creek formation (late upper Miocene) Marsland formation (early upper Miocene)
	Arikaree	Harrison formation (late middle Miocene) Monroe Creek formation (early middle Miocene)
		Gering formation (lower Miocene)
OLIGOCENE	White River	Brule clay: *Leptauchenia* beds; *Oreodon* beds
		Chadron formation or *Titanotherium* beds

has been described and mapped by Schlaikjer (1935). A large portion of this area is mapped as "Lower Harrison". Schlaikjer states:

> "This formation constitutes the high upland table lands of the area and is exposed principally on Sixty-Six Mountain, Bear Creek Mountain, in the southwestern corner of the area and in the vicinity of Deer and Six Mile creeks south of old Fort Laramie. Its maximum thickness is approximately four hundred feet, and it is mainly composed of light to dark gray unconsolidated sand and sandstones in the form of channels and concretions. . . . The lower Harrison formation rests with disconformity and unconformity on the Brule, and its lowest limits are generally marked by channel deposits".

Schlaikjer states at other places that "a duplicate sequence occurs" at Eagle's Nest and at Signal Butte and that at these points in Nebraska "the lowermost Harrison is composed of rather well stratified sands with only very thin sandstone layers".

Later (p. 118) he discounts the Gering formation, stating,

"I have inspected the Gering formation at the type locality south of Gering, Nebraska, and have made a preliminary study of it from that locality westward to Wyoming. It appears to be nothing more than an exaggerated eastward extension of the same conditions that prevail at the base of the lower Harrison in the Goshen Hole Area. In other words, the Gering represents only a local phase of the lower Harrison . . .".

It is not clear just what is meant by "an exaggerated eastward extension of the same conditions. . . ." The Gering formation is now considered the basal part of the Arikaree group.

Schlaikjer does not explain what became of the Monroe Creek formation in the Goshen Hole Area, but his descriptions, recorded sections, and the personal inspection by the writer and others in the field clearly indicate that Schlaikjer's "Lower Harrison" is in reality the Gering formation next above the unconformable contact on the Brule and the Monroe Creek formation in nearly the whole area. Probably the beds, which he calls "Upper Harrison" (now Marsland), "at the heads of Deer and Cherry creeks" are in reality true Harrison (old lower Harrison). The writer has personally examined the exposures at the heads of Deer and Cherry creeks, and Harrison beds are probably present here even though the lithology of the entire 400 feet or more of exposed section in this vicinity is not typical of all the zones which are probably present. There is some paleontological evidence of the presence of somewhat higher beds here than seem to occur generally in other parts of the area. On the other hand, there is also a possibility that faunas lived in the Goshen Hole Area earlier than almost identical faunas found in higher (?) beds in certain other areas farther to the northeast. Geographic distribution, which was controlled by general climatic conditions and water supply, has no doubt greatly affected the distribution of faunas in time almost as much as in geography. Continuity of significant lithologic zones is probably just as safe a criterion of contemporaneity of sedimentation as identity of fossils. The Gering and Monroe Creek formations with perhaps some Harrison formation are distinct lithologic zones and appear to extend into the Goshen Hole Area. It seems more logical to accept the continuity of the lithology as indicative of formational identity than to assume rather extreme facies change in lithology on the basis of paleontological evidence which may questionably indicate a somewhat later age for the beds.

Both Signal Butte and Eagle's Nest, points in Nebraska noted above as referred to by Schlaikjer, are capped with only the Gering formation above the Brule clay. The east end of Sixty-Six Mountain, which extends into Scotts Bluff and Banner counties, Nebraska, has Gering and Monroe Creek formations only above the Brule but does not have any Harrison formation on it, at least not in Nebraska. Remnant patches of Harrison (old lower Harrison) formation occur in some places on Wild Cat Ridge

farther east in Scotts Bluff, Banner, and southern Morrill counties; and a little wedge of Harrison, up to 20 to 40 feet thick, occurs at the Wyoming-Nebraska state line in western Banner County, where it extends nearly across T. 18 N., R. 58 W., in the bluffs between the Monroe Creek and Ash Hollow (Ogallala) formations.

Schlaikjer nowhere indicates that he intended to redefine the Harrison or to discard the Monroe Creek. He apparently thought that the Monroe Creek either did not occur in his area or that it correlated with and was included in the lower Harrison formation. He does not make this clear. In any case, a conformable Harrison (lower Harrison) over Gering would be anomalous, since in ascending order the Gering, Monroe Creek, and Harrison formations lie superimposed in the Pine Ridge and elsewhere. They are less easily distinguished toward the east along the Pine Ridge, that is, from Chadron eastward across northern Sheridan County, than they are westward across Dawes and northern Sioux counties. To the east and in South Dakota these formations correlate with the Rosebud.

The upper part of the Monroe Creek formation, about 100 to 110 feet thick, is exposed about 4 miles east of Agate where it is brought high in the structure of the Agate anticline. Here, it is overlain by the typical Harrison (lower Harrison) formation, up to 50 feet or more thick at the location of the old Agate test well for oil. Schlaikjer states that he has studied the lower Miocene "from the Rosebud-Pine Ridge district of South Dakota southwestward across Nebraska, via Chadron and Agate, into the Scotts Bluff-Goshen Hole areas. . . ." Remembering that the Harrison formation is thin or "pinched out" and in some places absent altogether east of Chadron, Nebraska (see Pl. 1), the writer believes that Schlaikjer either did not distinguish the Harrison (lower Harrison) from the Monroe Creek in the Pine Ridge or that he mistook the Monroe Creek formation for the Harrison. He might then easily have thought that the exposed 100 feet or more of Monroe Creek formation at Agate was the Harrison (lower Harrison) and that the overlying beds belonged naturally to the upper Harrison (now Marsland). As a matter of fact, there are no occurrences of Marsland (upper Harrison) at Agate. Some remnants of the Marsland formation do occur to the southeast and others to the northeast of Agate. A mistaken identity of formations as stated above might easily occur. It would then have been most natural for Schlaikjer to identify the Monroe Creek-Gering part of his Goshen Hole exposures as lower Harrison (now Harrison), and the occurrences of

> "light to dark gray sands and sandstones which contain some lenticular beds of clay and which, at certain levels are made up of more than fifty per cent of plant remains",

at the heads of Deer and Cherry creeks, as upper Harrison (now Marsland). In any case, Schlaikjer (1935) described and mapped the Gering and Monroe Creek formations as "lower Harrison", and his "upper Harrison" probably is true Harrison or old "lower Harrison". Gering, Monroe Creek, and remnant patches of Harrison formation occur in the "Goshen Hole Area" in just about the same way as they occur to the east in Nebraska, where, however, they are less deformed by structure.

OBSOLETE NAMES

General statement.—Certain obsolete names for "beds" or formations have attained some use, and a number of them have served a useful purpose; in some cases, their usage has been abused, and, because of too general or too divergent application to widely separated "beds", either stratigraphically or geographically, they have had to be abandoned. Four such outstanding Nebraska names are Loup Fork, Nebraska, Rosebud, and Republican River. The terms Loup Fork and Loup River have been confused and often used interchangeably; "Loup River" has been considered the higher and younger division. Both names are now abandoned by the Nebraska Geological Survey.

"Loup River"-"Loup Fork".—Simpson (1933) has summarized the "Loup River"-"Loup Fork" as follows:

> "Loup River, Loup Fork: Obsolete terms once in general use for part or all of the Miocene. The history of these unfortunate names is summed up by Osborn and Mathew (1909). The 'Loup River' was defined by Meek and Hayden in Nebraska as a formation, in largest part probably of Pleistocene age, but also made to include all the adjacent deposits down to the White River. Leidy continued to confuse the true or type 'Loup River', Pleistocene, with the much older Upper Miocene or Lower Pliocene ('Nebraska'). From this the name was generalized by Cope, in the form 'Loup Fork' as an epoch name in which he placed the Santa Fé. The name continued for a time in general use and the most diverse horizons were referred to it, from what is now considered Lower Miocene to the Lower Pliocene. It has no historic authenticity, unless 'Loup River' be retained for a Pleistocene horizon, and no accurate and exact meaning, so that it has necessarily been discarded for the most part, although occasionally employed for the Nebraska or Valentine stage".

"Loup River" has not been retained or resurrected for any part of the Pleistocene of Nebraska (Lugn, 1935, p. 127, 188-195). Osborn and Matthew (1909), noted above by Simpson, indicate that " 'Loup Fork' of Leidy, Marsh, also of Scott and Osborn . . . " applied mainly to the "Ogallala formation", also to the "Ogallala and Arikaree of Darton (in part)", and that it was equivalent in part to other formations in other areas, for example: Deep River, Pawnee Creek, Mascall, Santa Fé, Clarendon beds, Madison Valley beds, Archer, Rattlesnake, and Blanco; and to the Pontian and Eppelsheim of Europe. Osborn and Matthew also discuss the use of these terms by "Meek and Hayden, 1861-62", by Hayden, and by Hayden and Leidy.

"Loup River" is the original name and "Loup Fork" was its synonym for many years. Hayden (1869) and his associates applied the name "Loup River beds" to the part of the "Tertiary Rocks of Nebraska" which they regarded as "Pliocene". However, their "Pliocene" included everything now in the Ogallala and Arikaree groups down to the White River. This was also the lower division of Hayden's "bed F" (about equal to Ogallala and perhaps certain Pleistocene deposits in northern Nebraska) and "bed E" (in the main the present Arikaree group). The upper part of Hayden's "bed F" is Pleistocene loess (Loveland and Peorian), "most fully developed along the Missouri River from the mouth of the Niobrara to St. Joseph. Also in the Platte Valley and on the Loup Fork".

Nebraska beds of Scott.—Scott (1893) introduced the name "Nebraska" or "Nebraska beds" for the middle part of the "Loup Fork" as he understood it in 1893. He states:

> "The great Loup Fork formation has been much misunderstood, partly because it has not always been distinguished from the much later *Equus* beds, and partly because the separate horizons into which it may be divided have but lately been worked out. Of these there are three. The oldest is the Deep River horizon of Montana (*Ticholeptus* bed of Cope) and is characterized especially by a number of aberrant oreodonts and some peculiar genera of horses. . . . The second or middle horizon of the Loup Fork covers a vast area from Nebraska to Mexico, and has yielded a very large number of mammals. Of these the most characteristic is, perhaps, the antelope-like genus *Cosoryx*. An appropriate geographic name for this subdivision would be 'Nebraska'. . . . The latest of the three horizons of the Loup Fork may be called the Palo Duro, ——".

He thought that the Deep River and Nebraska horizons "belong unquestionably to the Miocene". Simpson (1933, p. 103) says of the Nebraska beds:

> "*Nebraska:* Proposed by Scott (1894A) for deposits of western Nebraska apparently synchronous with those now called Valentine (which see). Hatcher (1902) and Peterson (1906) applied the name to much older beds now included in the Upper Harrison.[4] The name is not in current use in either sense. ('Nebraskan' is a stage of the Pleistocene.)"

Rosebud.—Rosebud is the name applied to about 500 feet of pinkish sandstone and silty clay, the "Lower Miocene of southern South Dakota", and it has been used to some extent for the bluff-forming layers of the eastern part of the Pine Ridge, mainly in northern Sheridan County, Nebraska. The name was applied by Matthew and Gidley (1904). Matthew tentatively correlated the Rosebud with the lower Miocene of western Nebraska, that is, with the Harrison. In a general way, this is correct. However, all the formations and horizons of the Pine Ridge farther west in northern Sioux and Dawes counties have now been traced into the Rosebud of northern Sheridan County. The Rosebud in north-

[4] Upper Harrison, as explained, is now the Marsland formation.

ern Sheridan County encompasses all that is present of the following formations: Gering, Monroe Creek, Harrison (old lower Harrison), and at least the lower part of the Marsland formation (old upper Harrison). The Harrison (old lower Harrison) is nowhere in this area present in its full development. Excluding small remnant lenses, it was largely removed by erosion over the anticlinal arch which extends from the vicinity of the northwestern corner of Sheridan County southeastward near Hay Springs and across the State perhaps to connect up with the "Cambridge anticline" in southern Nebraska. The age of this folding and erosion seems to be post-Harrison (old lower Harrison)—pre-Marsland (old upper Harrison).

The Gering and Monroe Creek formations seem almost fully developed across northern Sheridan County except at the extreme east where both formations terminate. However, the Gering is indistinguishable from the Monroe Creek formation in this area on the basis of lithology alone, and the Gering is mapped with the Monroe Creek on Plate 1. The Gering horizon is known to be present because of the occurrence of "Gering" vertebrate fossils. Because of the erosion and "pinching out" of the Harrison in this area, the Marsland formation rests in contact with the upper part of the Monroe Creek in many places. (*See* Plate 1.)

Therefore, in so far as recognized in Nebraska (northern Sheridan County), the Rosebud buff and pinkish silty sandstones and clays correlate with the Gering, Monroe Creek, remnants of Harrison, and the lower part of the Marsland formation. This is not contrary to the opinions of others (Cook, 1915; Cook and Cook, 1933), but in most other discussions the question of correlation generally is narrowed down to a correlation of faunas and almost never includes the main formational units. Nevertheless, Perisho and Visher (1912) have indicated almost exactly the same correlation for the Rosebud as given above. O'Harra (1920) also has written of the Rosebud and other Tertiary formations of South Dakota and has indicated the same correlation for the Rosebud.

The term Rosebud is more or less superfluous in Nebraska. It has never been widely used within the State, and, in so far as Nebraska is concerned, it may be considered obsolete. However, the writer would not censure the retention of "Rosebud" in South Dakota and northern Nebraska as an inclusive term for the correlative formations where they are too indistinct for differentiation. It represents a facies of these formations in a certain area where finer differentiations are difficult.

Republican River.—Regarding Republican River, Simpson (1933, p. 107) states:

"*Republican River:* Lower Pliocene of northwestern Kansas, and, by extension, the corresponding stage of the Pliocene generally. The beds were extensively ex-

plored, especially for Marsh, and a large fauna is known, but there is no recent faunal or stratigraphic revision aside from such syntheses as that of Osborn and Matthew (1909) or references in describing individual fossils or faunas of similar age. It was generally placed in the 'Loup Fork', 'Upper Loup Fork', or 'Ogallala', but the more definite local name has been consistently applied by Matthew and is now generally accepted. Matthew (in Osborn and Matthew 1909) tentatively considered it as slightly later than the 'Nebraska' (Valentine), later reaffirming this more positively (1924B). Most recently, Matthew and Stirton (1930B) have, in passing, suggested that the Republican River may be in part composite, with the beds on Sappa Creek and elsewhere to the southeast later than those on Driftwood Creek and to the north. They continue to place the fauna, as a whole, as younger than Valentine, adding that it seems to be slightly older than their Goodnight-Hemphill (and hence equivalent to Clarendon)".

The exposures generally referred to the Republican River beds in southwestern Nebraska and adjoining portions of Kansas are now known to belong to the lower part of the Ogallala group, that is, they are equivalent to the upper part of the Valentine formation and the lower part of the Ash Hollow formation. The Republican River is neither a lithologic nor a stratigraphic unit, and the fauna, so called, is a "composite" and "mixed" one. The "Republican River" exposures should be referred to their proper formational horizons, and "Republican River" has been abandoned by the Nebraska Geological Survey.

As noted by Simpson above, Matthew and Stirton considered the fauna "as younger than the Valentine". This is in all probability correct, as the fossil-bearing horizon on Driftwood Creek, southwest of McCook, Nebraska, seems to correlate almost exactly with the Burge channel horizon at the top of the Valentine formation in northern Nebraska. The fossil-bearing channel deposits at Driftwood Creek also occur below the "caprock bed" and the *Krynitzkia* fossil seed zone, the same position as occupied by the Burge channel member of the Valentine. The "Valentine" fauna occurs in some places as much as 100 feet or more below the Burge level. The "beds on Sappa Creek" and other places are mostly higher and later than the fossil-bearing horizon on Driftwood Creek, and these higher horizons belong to the lower part of the Ash Hollow formation, the *Krynitzkia* and lower part of the *Biorbia* seed zones, all definitely higher stratigraphically than the level of the Valentine fauna at Valentine, Nebraska.

Considering all available evidence, it seems certain that the "Republican River" represents a composite of lithologic and faunal horizons and should be abandoned as a stratigraphic term. Furthermore, many of the fossil finds were obtained from the unconsolidated sands and gravels in pits under the lower slopes of the valleys west and southwest of McCook, Nebraska. Some fossils have been collected from higher beds, from the "mortar beds" along the sides of the valleys. The sand and gravels belong to the upper part of the Valentine formation, and the "mortar beds" levels contain the *Krynitzkia* fossil seed zone and the

lowermost part of the *Biorbia* fossil seed zone, all in the lower part of the Ash Hollow formation. Therefore, the fauna is much mixed, and also the described and published sections do not include the lower (Valentine) beds in the exposures. Thus it seems that fossils from as low as the Burge member or even lower in the Valentine formation have been credited as coming from "mortar beds" levels in the lower part of the Ash Hollow formation, all as "Republican River".

"Republican River" could not be redefined in any way to include all of the Ogallala group; it is not synonymous with Ogallala and could not be substituted for Ogallala.

CONCLUSIONS

The writer believes that the classification of the Tertiary system in Nebraska presented here is a workable classification, subject, however, to revisions as future more detailed and more exact knowledge may require; that the stratigraphic units now acceptable to the Nebraska State Geological Survey and employed by it for mapping are satisfactory and as generally acceptable under present circumstances as could be expected; also that the nomenclature presented here very properly may be applied for purposes of stratigraphic description and mapping to most of the Tertiary deposits of the High Plains. It seems probable that the use of fossil seed zones may well mark a new epoch in Tertiary stratigraphic investigation for nearly all late Miocene and Pliocene sediments in the Great Plains Region. The application of "stratigraphic methods" to the interpretation of Tertiary sedimentation and the age relations of deposits and contained faunas contributes much to the more accurate determination of faunal successions and animal evolution.

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University of Nebraska, Lincoln, Nebraska.
Manuscript received by the Secretary of the Society, December 24, 1938.
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1277-1310, 1 PL., 4 FIGS. AUGUST 1, 1939

STRUCTURE OF THE WASATCH-GREAT BASIN REGION

BY A. J. EARDLEY

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ABSTRACT

The major structural elements are of three kinds—folded and thrust belts, broad uplifts, and fairly resistant buttresses. The Colorado plateau and Wyoming basin are

the largest resistant elements. Along the western margin of the Wyoming basin lies the Idaho-Wyoming arc of folded and thrust strata. Along the western margin of the Colorado plateau is another north—south-trending group of folds and thrusts. The Colorado plateau and the Wyoming basin are separated by the broad east—west-trending Uinta uplift. A small buttress of recurring positive character in northern Utah lies directly in the path of the southwesterly-trending Idaho-Wyoming arc folds which are deflected around it. The northward-trending folds south of the Uinta uplift veer westward, unite with the Idaho-Wyoming arc folds, and continue in the southwest trend of the arc past the Sheeprock uplift. The Cottonwood uplift crowds the strata into folds and small thrusts against the west end of the Unita uplift and the southeast side of the northern Utah highland.

The evident lack of alignment of the Laramide folds and thrusts with the Basin and Range faults is most striking, and apparently a Laramide control is only local. The foliation of the pre-Cambrian rocks, inherited from pre-Cambrian time, seems to have exerted little influence on the course of the Laramide folds or the Basin and Range faults, although the northern Utah highland as a mass conspicuously deflected the Laramide folds around its margins. The highland did not affect the Basin and Range faults in any discernible way.

INTRODUCTION

PURPOSE OF INVESTIGATION

The original purpose of the investigation was to decipher the structure of the Uinta-Wasatch junction, the effects of the pre-Cambrian structures on the later structural trends, and the interrelation of the pre-Cambrian, Laramide, and Basin and Range deformations. During the course of the work other problems developed, such as the structural significance of the basins of sedimentation, the nature and amount of crustal shortening, and the relation of the Laramide structures in north-central Utah to those in Idaho, Wyoming, and Colorado. Little attempt was made to summarize the succession of orogenic events during the Laramide revolution, and no attempt was made to date them. This phase of the structural history is being treated by Spieker (1939?).

SOURCES OF STRUCTURAL DATA

The literature on north-central Utah was reviewed, and an index map prepared (Fig. 1) showing the local areas covered by detailed or reconnaissance structural maps. Fifteen different areas within the region have been mapped. They represent about 50 per cent of the exposed bedrock surface. The authors of the local maps and the date of publication are printed within their respective areas on the index map, and these names and dates are guides to the complete references listed in the end-bibliography. The remaining 50 per cent (horizontally lined areas) of bedrock surface was covered by the writer only sufficiently to trace the major structural trends from one mapped area to another. Base maps for this reconnaissance work were usually topographic maps of the King and Powell surveys or United States Forest Service maps.

The portion of the tectonic map of the United States in preparation by T. S. Lovering was used in compiling a simplified version of the

structures in Idaho, Wyoming, and Colorado, adjacent to the area under consideration. (*See* Figure 3.)

ACKNOWLEDGMENTS

The writer is indebted to Professor T. S. Lovering, of the University of Michigan, for advice on the graphic representation and interpretation of the structural trends and for critical reading of the manuscript. Professor E. M. Spieker, of the Ohio State University, has also read the manuscript and has offered valuable suggestions regarding the Laramide revolution.

The cost of field work was partially defrayed by grants from the Faculty Research Fund and the Horace H. Rackham School of Graduate Studies Research Fund of the University of Michigan.

LARAMIDE TRENDS

DEFINITION

By "Laramide trends" is meant the folds, faults, and intrusions of the Laramide revolution. The folds range from gentle open anticlines and synclines to those that are close and overturned. The faults are either vertical, high-angle reverse, or low-angle thrusts. The intrusions are chiefly small stocks with both discordant and concordant contacts. The Laramide structures in the Wasatch Great Basin region began to form in post-Jurassic—pre-Upper Cretaceous time, and several unconformities mark the repetition of deformation until the Uinta epoch of the Eocene (Spieker, 1938, p. 2040; 1939?; Spieker and Schoff, 1936, p. 104; Eardley, 1939?) and, perhaps, even later.

If the Laramide revolution began in early Cretaceous time, the closing phases of the Sierran revolution were probably still in progress. Such a broad usage of the term Laramide revolution is, perhaps, not desirable, but, since the subject is being treated comprehensively by Spieker (1939?), it is not advisable at this time to restrict the term to a shorter interval of time.

CORRELATION OF TREND LINES

After the trend lines had been drawn in the areas for which structural data were available, including those obtained in reconnaissance by the writer, a number of connections between ranges were immediately apparent. For instance, the two anticlinal crests and the intervening synclinal trough of the Tintic district (Lindgren and Loughlin, 1919, Pl. 1) lead directly to the anticlines and synclines of the southern half of the Oquirrh Range (Gilluly, 1932, Pl. 12). (*See* Plate 1 and Figure 1.) The exact position of the syncline and anticline in the intervening valley, Cedar Valley (T.1S., R.2W.), is not known but cannot be far from

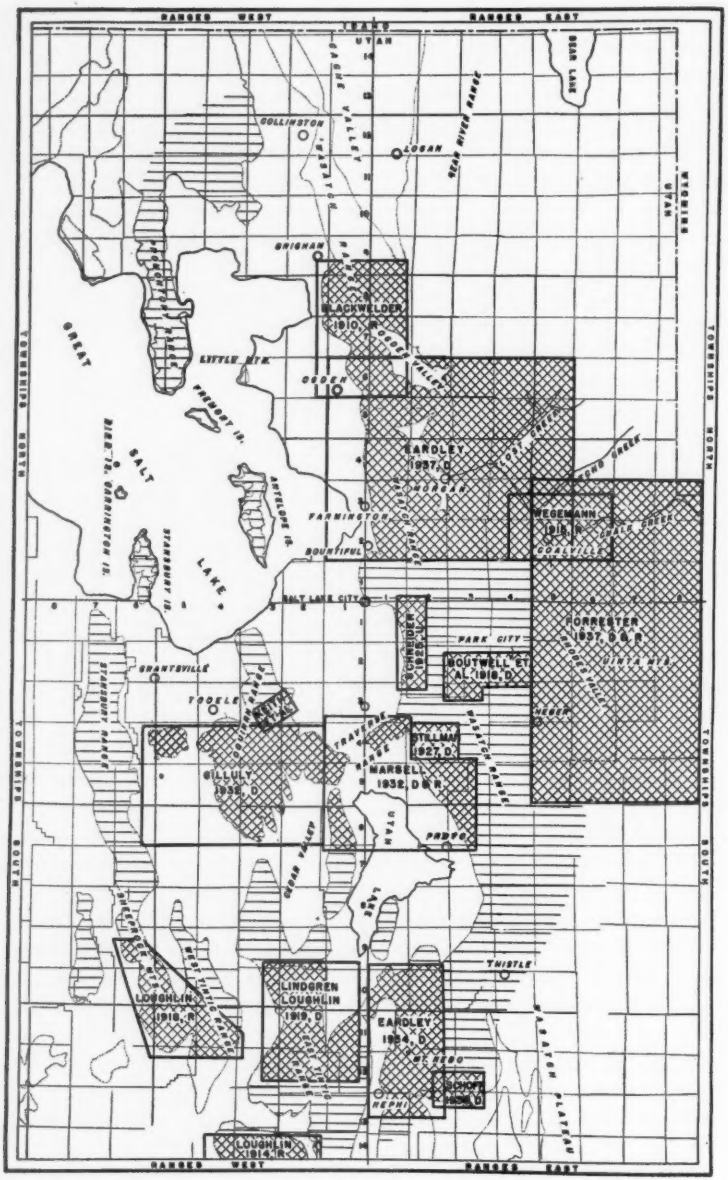


FIGURE 1.—*Index map of the north-central Utah area*

D, R, and U respectively represent detailed map, reconnaissance map, and unmapped but with cross section. The horizontal ruling indicates the areas covered in reconnaissance by the writer.

positions arrived at arbitrarily by uniform spacing. The anticline of Stansbury Island (T.1N., R.6W.) is in direct alignment with the anticline in the north end of Stansbury Range, and, in both, the same sedimentary formations are exposed.

Guided by these more obvious correlations, additional but more speculative connections were drawn. The area of pre-Cambrian crystalline rock was outlined by projecting its contacts in the Bountiful-Brigham pre-Cambrian part of the Wasatch Range westward in conformity with the adjacent structural trends so as to include the pre-Cambrian rock islands of Great Salt Lake. Near the south end of the Bountiful-Brigham area Cambrian (?) limestone exposures of the mountain front at Beck's Hot Springs are close to the crystalline rocks in the mouth of Mill Creek Canyon (Mueller's Park), and the beds exposed nearest the contact dip gently northward toward the crystalline rocks (R. E. Marsell, oral communication). Although the Wasatch conglomerate overlies the contact in the critical location, the Cambrian quartzite appears to be cut out at the mountain front. Near the head of Mill Creek Canyon it is exceedingly brecciated, dips steeply, and strikes southwestward toward the Cambrian (?) limestone area of the mountain front. From this structural setting the writer postulates a thrust fault of small displacement that trends east–west and bounds the southern margin of the crystalline complex. The north end of the Bountiful-Brigham crystalline area is broken by the slices of the Willard thrust fault (Blackwelder, 1910). The writer has reworked this area and interprets Blackwelder's east–west transverse faults as sharply curving branches from the main northeastward-dipping thrust faults. Northward the branch faults veer toward the west and have influenced the interpretation of the position of the crystalline rock contact under the valley alluvium at Brigham (T.8 and 9N., R.1W.).

Late pre-Cambrian rocks on Promontory Point strike slightly north of east and dip north. Although remnants of the basal members rest on the crystalline rocks of Antelope Island, the line separating the thick sedimentary beds from the crystalline rocks probably runs somewhere between Promontory Point and Fremont Island. The projection of the anticline southward under the Bear River embayment toward Little Mountain is chiefly interpretative and, in harmony with this interpretation, the writer has swung the pre-Cambrian contact several miles north of Little Mountain.

Bird Island is composed of tillite which the writer has correlated tentatively with the basal beds of the late pre-Cambrian. It is believed, therefore, that the crystalline complex is near the surface just east of the island. Carrington Island, 4 miles south of Bird Island, is

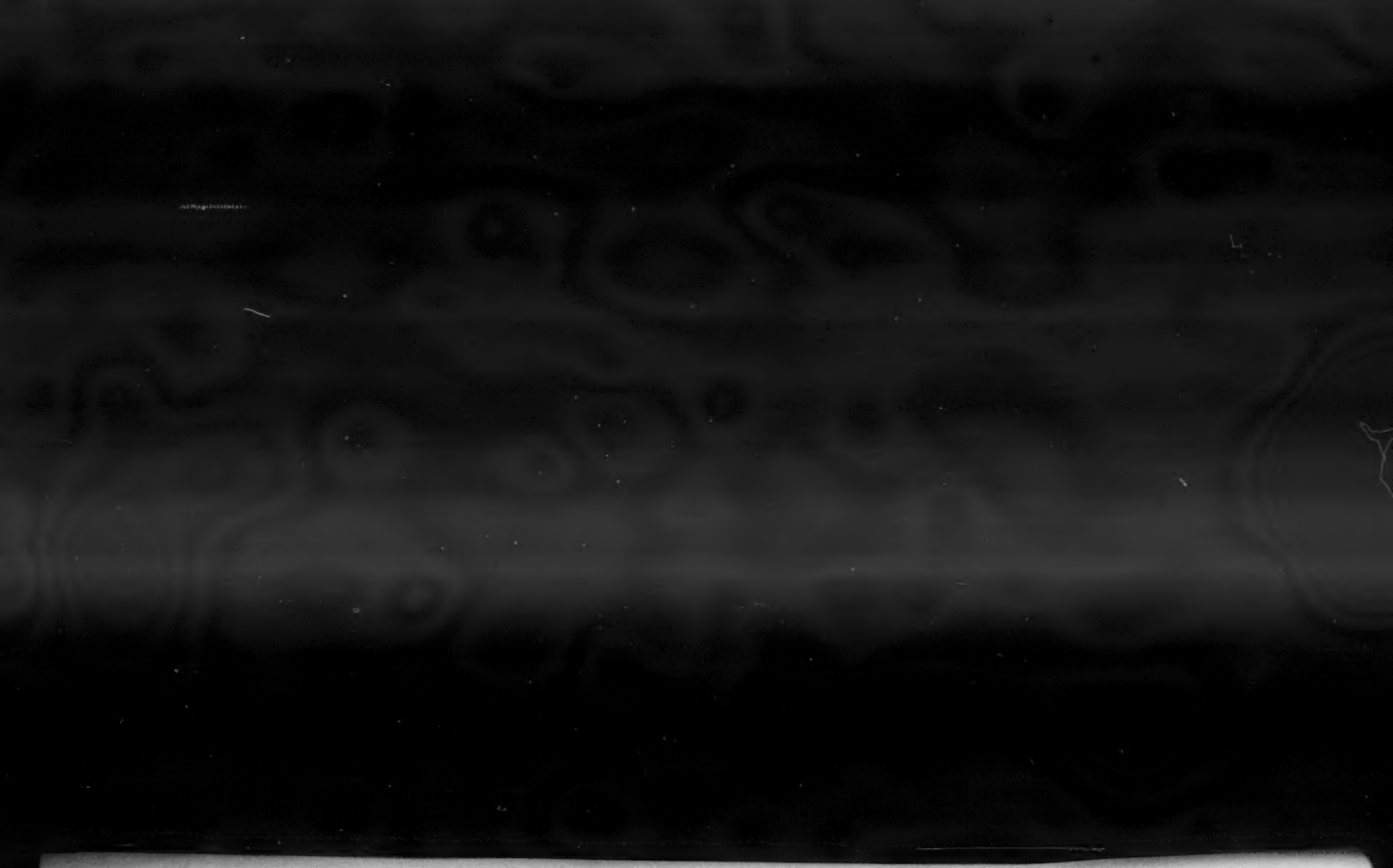
composed chiefly of limestone, calcareous shale, and quartzite of Mississippian or pre-Mississippian and post-Cambrian age. For this reason the approximate contact of the crystalline rocks with the thick Paleozoic veneer between the two islands is rather sharply defined, and the structure is probably a fault or a sharp upturn of the beds. The structural trends of Stansbury Island and the north end of the Oquirrh Range have guided the drawing of the contact from Carrington Island to the Wasatch front.

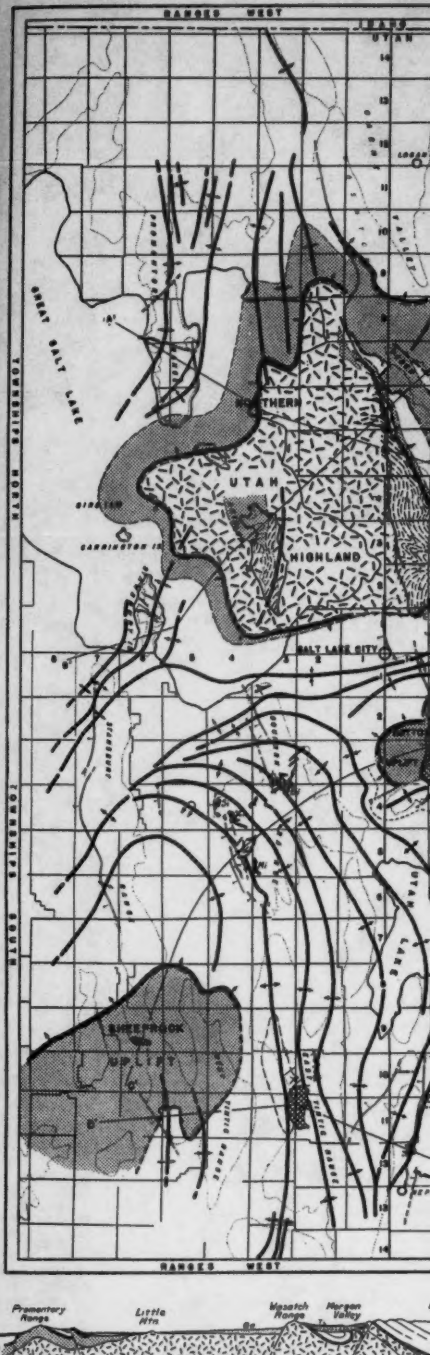
The north end of Antelope Island is composed of Cambrian(?) quartzite (Pl. 1, sec. B′-B), and it is probable that other remnants of Paleozoic rock occur under the valley alluvium on the pre-Cambrian crystalline complex outlined above. It is thought that these remnants are thin, for the area was dominantly a highland during late pre-Cambrian and Cambrian time and afterward suffered much erosion. This aspect of the pre-Cambrian crystalline area will be discussed under the heading "Northern Utah Highland."

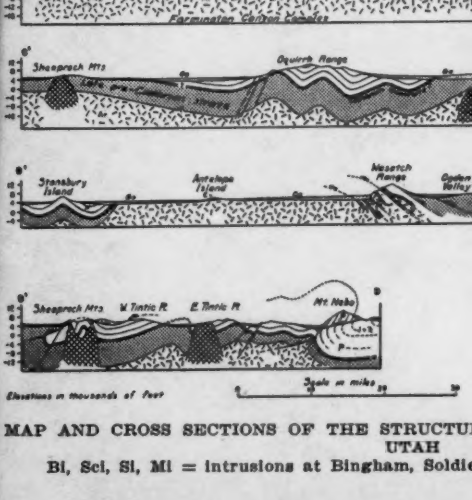
STRUCTURAL ELEMENTS

Uinta axis.—The Uinta axis is a salient feature of the structural pattern of the Wasatch-Great Basin area. It is marked by three uplifts in which late pre-Cambrian quartzites, shales, slates, and phyllites are exposed. The easternmost of the uplifts is that of the Uinta Mountains which is, in general, a broad elongate arch involving approximately 35,000 feet of sediments. The strata of the flanks are steeply flexed and in places ruptured by steep normal and reverse faults. The beds (mostly late pre-Cambrian) between the lateral flexures and faults are nearly flat, rising gently toward the topographic crest of the range (Powell, 1876, Pl. 3; Forrester, 1937, Pl. 3). (*See* Plate 1, cross section A′-A.) Farther west in the central Wasatch (T.2S., R.2E.) a second uplift occurs where about 20,000 feet of strata has been arched into a sharp dome. Cambrian quartzite and late pre-Cambrian quartzites, argillites, and shales 12,000 feet thick constitute the lower part of this series. The underlying crystalline rocks of the Farmington Canyon complex crop out in the mouth of Little Cottonwood Canyon (Schneider, 1925). (*See* Plate 1, cross section C′-C.) The beds of the Cottonwood uplift have been intruded by stocks of granodiorite. The largest outcrop—the Cottonwood intrusion—is centered slightly south of the apex of the dome, and the others—the Park City intrusions—extend eastward along a projection of the Uinta axis on the east flank of the dome and across an anticline and two synclines in Paleozoic and lower Mesozoic strata. The intrusions cut complicated thrust faults of limited north–south extent and relatively small displacement.

The third uplift along the southwestward projection of the Uinta axis centers in the Sheeprock Mountains. The late pre-Cambrian quartzites

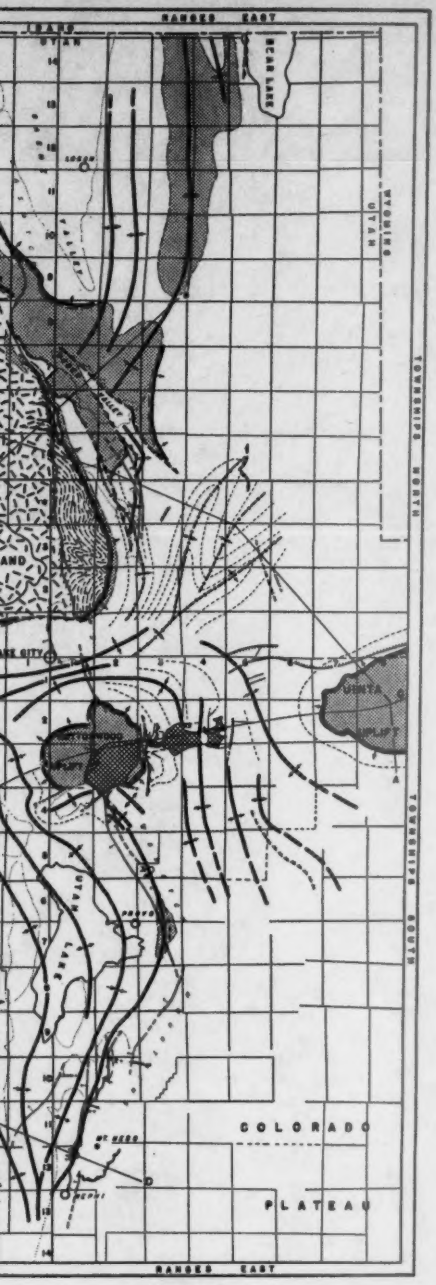


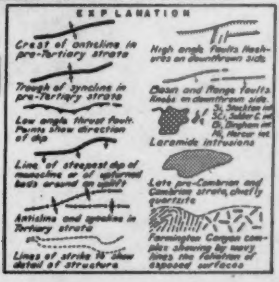




MAP AND CROSS SECTIONS OF THE STRUCTU
UTAH

Bi, Sci, Si, Mi = intrusions at Bingham, Soldie




EXPLANATION
Crest of anticline in pre-Tertiary strata
Trough of syncline in pre-Tertiary strata
Low angle thrust fault. Points show direction of dip
Line of steepest dip of monocline or of upturned beds around an uplift
Anticline and syncline in Tertiary strata
Lines of strike to show detail of structure
High angle faults. Hachures on downthrown side
Basin and Range faults. Knobs on downthrown side
Laramide intrusions
Late pre-Cambrian and Cambrian strata, chiefly quartzite
Farmington Canyon complex showing by wavy lines the foliation of exposed surfaces

STRUCTURAL TRENDS OF NORTH-CENTRAL UTAH

am, Soldier Canyon, Stockton, and Mercur.

crop out in the West Tintic Range, in the Sheeprock Mountains, and in the range to the west—the Simpson Mountains. (*See* Plate 1 and Figure 1.) On the northwest flank the Sheeprock uplift appears fairly regular, but to the south the strike and the dip readings taken on a hasty survey suggest complicated folds, whose trends were not determined. The structure is complex also in the West Tintic district (T.11S., R.5, and 6W.), where Loughlin (1918, Fig. 46) has described Algonkian strata thrust over folded Paleozoic strata. It is evident that the Sheeprock uplift is structurally far more complicated than the Uinta Mountains.

In the same manner that the Cottonwood uplift is separated from the west end of the Uinta uplift by a small synclinorium, so is the Sheeprock uplift separated from the Cottonwood. The horseshoe-shaped pattern of the folds around the Sheeprock uplift and between it and the Cottonwood uplift suggests a structural relation of the two. The Ophir dome lies on the line of a smooth curve drawn from the west end of the Uintas through the Cottonwood uplift to the Sheeprock uplift. (*See* Plate 1 and Figure 4.) The Bingham (Bi), Soldier Canyon (Sci), and Stockton (Si) intrusions lie to the north of this line, and the Mercur intrusion (Mi) to the south.

The problem of the westward continuation of the Uinta axis has been discussed in some detail by Gilluly (1932, p. 71-73). He considers the views of Butler (1918, p. 100-105), Olmstead (1921, p. 437), Spurr (1913, p. 475-485), Porter (1924, p. 650-651; 1927, p. 793-796), and Beeson (1927, p. 768-792) and makes the following conclusions:

"The supposed extension of the Uinta anticline to the west through Bingham and its superposition upon the folds of northwest trend is definitely not a fact, and it remains questionable whether or not the nearly east—west cross axis through Ophir is a reflection of the north—south compression to which the Uinta fold most probably owes its origin. The association of the intrusive of the Bingham district with a very deep syncline is in marked contrast to the anticlinal location of the Cottonwood and Park City stocks, and there is no consistent evidence to support the suggestion that the Uinta uplift is due mainly to igneous intrusion. The existence of igneous rocks from Mercur on the south through Ophir, Soldier Canyon, the Stockton district, and Bingham on the north renders an east—west trend of intrusions in the Ophir Range a matter of some doubt also."

Recently a paper by Billingsley and Locke (1939, Fig. 10) shows the Uinta axis to project apparently along the axis of the east–west anticline of Traverse Range (Marsell, 1932, Fig. 32). However, this anticline does not continue across the synclinorium folds of the Oquirrhs. The absence of a cross fold through the synclinorium and the failure of the Oquirrh Mountains intrusions to line up along the Cottonwood-Uinta axis projected across the Oquirrhs give convincing evidence that the "Uinta axis" does not there exist as a "fold axis." There is little doubt about the actual nature of the geologic structures, but Gilluly's use of the term "axis" may be questioned. The writer believes that the curved align-

ment of the three uplifts and the Ophir dome, and the emphasis imparted to this alignment by the ground plan arrangement of the folds, make it a major structural element. Whether this element be called a trend, a lineament, or an axis is a matter of preference because no name for a structure of this type has become standardized. Since "Uinta axis" has long been in the literature, the writer prefers to retain it, but without inferring that it is an axis of an anticlinal crossfold along which a series of intrusions are precisely aligned.

Folds and thrusts.—Great length is a main characteristic of the Laramide folds. The fold that passes through the Ophir dome has been traced continuously for 90 miles. Such folds, as typified in the Oquirrh Range, have flanks that dip 30 to 60 degrees. They involve principally Paleozoic strata. The anticline that extends from Nephi northward in the Wasatch Range is overturned and thrust in places toward the east (Schneider, 1925, Fig. 5; Eardley, 1933a, Fig. 9). In the Spanish Fork-Provo section the structure is probably even more complicated (A. A. Baker, oral communication). The upright anticline through the Ophir dome (Pl. 1 and Fig. 4) passes into an overturned fold in South Mountain, T.4S. and R.5W. (Gilluly, 1932, p. 12). Throughout most of their extent in and east of the Wasatch Mountains the Eocene formations have been only gently flexed but in a few places they are rather intensely folded. The area northeast of Salt Lake City is the only one for which a map of the Tertiary structures is available (Eardley, 1939?). Here the folds in the Wasatch conglomerate and associated formations appear approximately parallel with but not superposed on the folds in the underlying Mesozoic and Paleozoic strata. (*See* Plate 1 and Figure 3.)

The most conspicuous thrusts in the Wasatch Range lie northwest and southeast of Ogden Canyon (Blackwelder, 1910). They dip 10°–55° NE. and die out to the southeast where the structural trends of the central Wasatch and Uinta mountains run athwart their course. They apparently die out also to the northwest. There is no evidence that the Willard thrust connects with the Bannock thrust of southeastern Idaho or with the small thrusts in Big Cottonwood Canyon to the south. If the Bannock thrust does extend far south into Utah, it would lie probably 10 or 15 miles east of the Willard thrust. The only clue to its position that the writer knows is a possible thrust fault in Toone Canyon, a tributary of Lost Creek Canyon (T.5N., R.5E.). The Willard and associated thrusts appear to be local developments along the northeast side of a buttress called the northern Utah highland.

The Sheeprock thrust (Pl. 1, and Figs. 3, 4) has considerable horizontal displacement. It and the Nebo thrust are shown in Plate 1, cross

section D′-D. The Sheeprock thrust has not been traced to the southwest, and hence its significance is not clear although it appears to be the major structural element of the Sheeprock uplift.

A fault running to the northeast diagonally across Promontory Range is also probably a thrust, and, although no measurement of the dip of the fault plane could be taken, the minor folds indicate relative overthrusting to the northwest. The structure believed most probable is shown in Plate 1, section A′-A.

High-angle faults.—The areas mapped in detail, particularly the mining districts, are known to contain steep faults most of which trend at varying angles to the fold axes. The average trend of the high-angle faults is east–west, although many exceptions are known. The displacement on most of them is less than 500 feet, but on a few it is several times this amount. Marsell (1932, p. 88) calls these the "older" faults to distinguish them from the Basin and Range faults. Some of them are pre-mineral, and some are post-mineral; some were mineralized and suffered subsequent movement. Some were later than the main phases of the folding but may have been, perhaps, contemporaneous with the later phases of the folding. Some were earlier than the intrusions, and some were later (Gilluly, 1932, p. 74-90). All the faults of this group are, however, older than the Basin and Range faults but renewed displacement on some may have occurred during the Basin and Range orogeny. They are not the type of high-angle reverse or thrust faults that strike approximately parallel to the axes of the folds. The flank faults of the Uinta Mountain, which are more strictly longitudinal, do not belong in this class.

Some of the faults were caused by vertical forces that acted during folding or by a differential release of pressure after folding. The clusters of faults in areas of igneous rocks are probably related in origin to the igneous activity. Large transverse faults have been mapped, however, that are not clearly related either to the folds or to igneous activity. Some of these were discovered in the southern Wasatch Mountains (Eardley, 1933a, Map 11), and others in the Promontory Range and Stansbury Island. It is suggested that the faults in part are tensional structures normal to the direction of slight crustal elongation during the Laramide revolution. They are analogous, perhaps, to Cloos and Balk's cross joints which are at right angles to compressional folds. As several episodes of folding probably occurred, it is probable that there were several episodes of cross faulting.

Northern Utah highland.—A feature of the structural pattern equally as prominent as the Uinta axis is the buttress of pre-Cambrian crystalline

rock north and northwest of Salt Lake City. It stood as a highland through parts of late pre-Cambrian, Paleozoic, and Mesozoic time, and during the Laramide revolution it acted as a small semirigid shield in the midst of an area of failing sedimentary rocks.

The northern Utah highland consists principally of early or middle pre-Cambrian gneisses and schists, which, previous to injection and metamorphism, were sedimentary rocks and sills or lava flows. The stratification of these rocks is still well preserved, and in places the beds are folded into structures simple enough to be mapped. An open anticline trends N. 60° W. through Bountiful Peak and is flanked on both sides by open synclines, all of which plunge about 15°–20° NW. The structures of the Bountiful Peak area are revealed better than those in other parts of the area. Across Weber Canyon the beds strike persistently northwest in harmony with the axes of the folds of the Bountiful Peak area. The new CCC road from Bountiful to Bountiful Peak (T.2N., R.1E.) passes over an area of extremely distorted beds where the structure was not deciphered. From Shepard Creek north to Holmes Creek (T.4N., R.1E.), across the mouths of the intervening canyons, there is a zone of distortion which may have originated from shearing in the Laramide revolution. On Antelope Island the general strike of the beds is north, except at the southern end of the island where the strike is northeast. Dips from 60 to 80 degrees prevail.

The Laramide folds follow the boundaries of the buttress closely. The anticline trending northwest through Stansbury Range (T.1 and 2.S., R.7W., Pl. 1 and Fig. 1) is split and deflected by the highland, one branch going to the east through the central Wasatch and the other to the north through Stansbury Island and then northeast into the buttress. The north–south folds and faults in Promontory Range are deflected to the southwest around the buttress. The strata are upturned around the entire eastern margin and are paralleled approximately by adjacent folds and low-angle faults. The Willard and associated thrusts parallel the northeast boundary. The conclusion is drawn that the pre-Cambrian crystalline rocks of the northern Utah highland acted as a small fairly resistant shield caught in the zone of compressional deformation. About its margins the sedimentary rocks buckled and sheared.

In the middle of the northern Utah highland on Antelope Island the Cambrian (?) quartzite and some basal beds of the late pre-Cambrian sequence occur in sedimentary contact on the crystalline rocks. The sedimentary beds strike N. 7° E. and dip 33° W. near the crest of the island but flatten appreciably toward the west side. Since the Cambrian beds on Antelope Island are folded and 15 miles eastward are overlain conformably by the Paleozoic and Mesozoic sequence, the folding appears

to be of Laramide age. The buttress of the northern Utah highland, therefore, probably did not stand unyielding to deformation during the Laramide revolution. An anticlinal axis just east of the crest of Antelope Island is suggested as a continuation of the axis from the north through Bear River Bay. (*See* Plate 1 and Figure 4.) The Stansbury Island anticline is believed to extend into the highland but probably dies out a short distance to the north. Although this picture of the deformed Farmington highland is incomplete and only approximate, it seems probable that the highland was more resistant to deformation than the surrounding crust and that it yielded by open folding while the adjacent crust was severely folded and sliced by thrusts, perhaps with ten times as much horizontal shortening. If the highland and the surrounding crust be considered as analogous to plastics, one with a viscosity higher than the other, then, if both were subjected to the same deforming stress, both would flow and deform, but for the same time the amount of yielding in each would be a function of the viscosity. Rather than view the northern Utah highland as an area of infinite viscosity that resisted the Laramide forces without deformation, it seems better to regard it as a section of the earth's crust in which the viscosity exceeded that of the surrounding crust.

The relationship of the open folds east of Logan and north of Ogden Valley to the Willard thrust sheets is especially interesting, for here the thrusts trend across the folds. The thrusts already have been described as locally developed structures that die out to the southeast and northwest. The rocks of the thrust area consist dominantly of the late pre-Cambrian and Cambrian metamorphosed sediments and were probably not so pliable as the younger strata. They yielded mainly by shearing.

It seems improbable that the folds and thrusts were formed at the same time. Both are compressional structures and apparently were caused by separate forces. Those causing the thrusts acted locally in a northeast–southwest direction, and those causing the folds in a northwest–southeast direction. The thrusts might have occurred first during the time of greatest compression. Less intense compressional forces thereafter from a direction nearly at right angles to the former could have resulted in failure by flexing. The thrust planes should then be folded along the axes of the folds. Mapping of the thrust planes has not been sufficiently detailed to enable the writer to determine positively whether this is true. The opposite sequence may have occurred with folding prior to thrust faulting. During the early stages of the Laramide revolution folds were formed first in the more flexible Paleozoic rocks under gentle horizontal pressure, and their trends were determined more by the Uinta uplift than by the buttress of the smaller northern Utah highland. After the forces

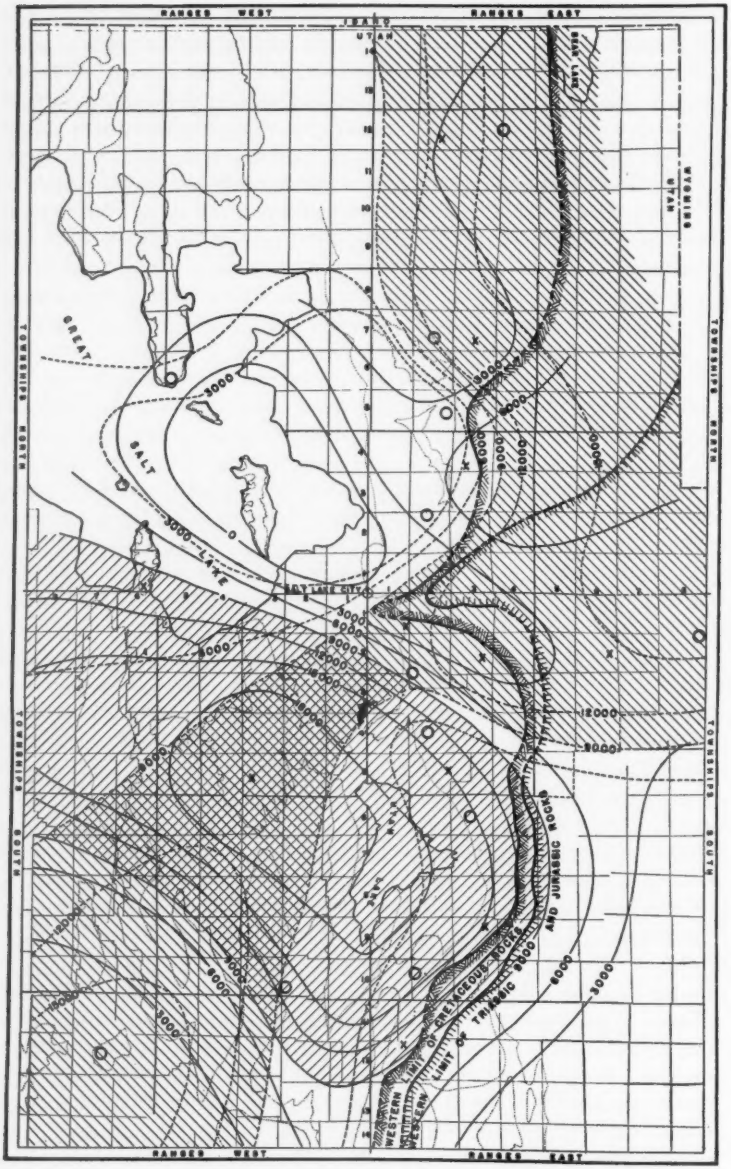


FIGURE 2.—*Isopach map of the thick Paleozoic and late pre-Cambrian sequences of strata in the Wasatch-Great Basin region*

The Mesozoic shorelines are also shown. Solid contours outline very approximately the Pennsylvanian basin, and the dashed contours the combined late pre-Cambrian and Cambrian basins. Small circles show localities where approximate thickness measurements are available for the late pre-Cambrian and Cambrian strata; crosses mark places where approximate thickness measurements are available for the Pennsylvania strata.

had become greater, the buttress was crowded intensely against the stratified rocks to the east, where they failed by shearing along quasi-imbricate planes parallel to the margin of the highland and across the fold trends. From a mechanical standpoint this theory seems more plausible. It is supported by the available structural evidence but still lacks confirmation from stratigraphic evidence.

The regional map (Fig. 3) shows that the northern Utah highland is in line with the northwest-trending Uncompahgre highland of Colorado and Utah. A positive area in central Nevada has been described by Nolan (1928, p. 153-161). It trends slightly east of north and came into existence during the late Paleozoic. Its rocks consist of a thick series of Cambrian and Ordovician strata. The northern Utah highland appears to have been a positive area as early as late pre-Cambrian. The relation of the northwest trend of the Uncompahgre highland and northern Utah highland (Fig. 3) to the central Nevada Paleozoic positive area is not known and remains an interesting subject for future investigation.

Late pre-Cambrian and Cambrian basin.—The observed Laramide folds are restricted chiefly to the areas of Paleozoic sedimentary rocks; the uplifts and thrusts chiefly to the areas of exposed late pre-Cambrian and Cambrian quartzites, phyllites, slates, and shales. The western margin of the Mesozoic strata south of Salt Lake City marks approximately the eastern limit of the pronounced folds.

An isopach map (Fig. 2) was made of the late pre-Cambrian and Cambrian beds in order to develop their relation to the areas of deformation and to the northern Utah highland. The data from the Wasatch Range suffice to show the detail desired, but not from the ranges to the west where considerable interpretation is necessary. No accurate data were available for the thick late pre-Cambrian section in the Sheeprock Mountains and vicinity, and the estimate of thickness made from the scattered dip readings taken by the writer is only approximate. It indicates that the late pre-Cambrian probably is more than 15,000 feet thick.

The isopach map shows the variation in thickness of the Cambrian and late pre-Cambrian quartzites, phyllites, and shales. It clearly defines an area where the sediments are very thin or absent. The thicknesses depicted are approximately the same as they were at the time of the Laramide revolution but in places they are less than the original total thickness of the late pre-Cambrian sediments plus that of the Cambrian quartzite. An angular unconformity between the two sequences truncates in places the entire 11,000 feet of Algonkian strata (Hintze, 1913, p. 98, fig. 3). The Cambrian quartzite rests directly on the schists and gneisses only along the northeastern, eastern, and southeastern boundaries of the

present exposures of the crystalline complex. The late pre-Cambrian rocks that are thrust over the Cambrian quartzite along the northeastern and eastern margins once lay several miles to the east where they separated the Cambrian rocks from the crystalline complex as a wedge thickening eastward.

Three alternative explanations of the origin of the unconformity occur to the writer: (1) The crystalline area defined by the isopach map was covered with 5,000 to 10,000 feet of late pre-Cambrian sediments (measured at Promontory Point and Big Cottonwood Canyon, respectively), and then, following elevation of the highland area, most of this thickness was removed by erosion. Subsidence then resulted in widespread deposition of Cambrian sand, which over the highland area rested partly upon the metamorphic complex. (2) The highland area stood in relief at the beginning of late pre-Cambrian sedimentation, and the late pre-Cambrian sediments were deposited as wedges that thinned against its shores. Later subsidence resulted in the widespread deposition of Cambrian sands over the highland and over the coastal and off-shore late pre-Cambrian sediments. (3) The highland subsided slightly and was buried by a thin veneer of sediments during the late pre-Cambrian. It was then elevated and stripped of its veneer in most places before the deposition of Cambrian sands. As the writer interprets the isopach data, the area of greatest subsidence during late pre-Cambrian sedimentation probably took the form of a , the stem of which extends from Salt Lake City southwestward past the Sheeprock Mountains, the left arm northward into Idaho along the site of the Wasatch Range, and the right arm eastward along the Uinta Mountains. The stem narrows perceptibly near the junction of the arms. In the Cottonwood district the sediments are composed chiefly of sand and pebbles. To the northwest on Promontory Point the series has thinned to less than half the Cottonwood section and is composed of finer material, silt, and clay in its lower half and sand in its upper half. The source of the sediments, therefore, appears to have lain to the southeast probably in the northern part of the Colorado Plateau. Although the sediments thin appreciably toward the northwest, they probably covered the highland area. Since more than 600 feet of sediments was deposited to the northeast in the Brigham City section of the Wasatch before the crystalline complex in Antelope Island was buried, it is probable that the highland stood in relief at the beginning of late pre-Cambrian sedimentation. Perhaps not more than 3000 feet of late pre-Cambrian sediments was ever deposited over the highland, and most of this must have been removed by erosion prior to the deposition of the Cambrian sands. This third theory is, in substance, a compromise between the first and the second, *viz.*, it postulates subsidence

or burial of the highland but only to the extent of about 3000 feet during late pre-Cambrian sedimentation, and then elevation and erosion, not of 11,000 feet of strata, but of about 3000 feet, before the deposition of the Cambrian sands. The highland area retained its "positive" character in contrast to the great submergence around it (the west side is not known) and then reasserted this tendency by rising and suffering erosion before burial by Cambrian sediments.

This old highland may have extended at times over 50 miles to the northwestward and may have joined with an exposed pre-Cambrian area in the Grouse Creek Range. Data are not sufficient, however, to determine this point.

The writer concurs with Spieker (1938, p. 2038) in questioning the interpretations of Forrester (1937, p. 626, and fig 1), who states that the "Algonkian rocks if omitted would not change the present geosynclinal picture." It appears to the writer that a very important element of the structural control has been the localized thickening of late pre-Cambrian beds whose distribution is in places quite different from that of the Paleozoic and Mesozoic rocks. The western extension of the Uinta axis follows the stem of the Y-shaped late pre-Cambrian basin and seems to have determined the position of some of the Laramide structures. Before further discussion, however, it is best to consider the extent of some other sedimentary rocks.

Pennsylvanian basin.—Gilluly (1932, p. 34-38) described a very thick series of Pennsylvanian beds in the Oquirrh Range, and the writer (Eardley, 1932, p. 322) subsequently noted a thickness in excess of 10,000 feet in the southern Wasatch Mountains. In addition to the thick Pennsylvanian series, Baker (oral communication) has recently discovered a comparable thickness of Permian beds in the Provo-Spanish Fork section of the Wasatch, which, as a structural control, probably should be considered together with the Pennsylvanian beds. The conspicuous open regular folds of the Oquirrh and Tintic ranges are principally confined to the area of Pennsylvania rocks. The Cambrian strata gradually thicken westward, but the Madison beds become somewhat thinner in that direction, and limestones of Brazer age overlap them, probably from the west and north. Devonian, Silurian, and Ordovician beds appear in the northern Wasatch and thicken northward into Idaho. Silurian and Ordovician beds thicken southward from the Tintic district. Their thickness relations are shown in Gilluly's diagram (1932, p. 39, fig 6). The total thickness of the pre-Pennsylvanian and post-Cambrian beds (post-Tintic or Brigham quartzite) varies relatively little in contrast to that of the Pennsylvanian. It appears, therefore, that the Pennsylvanian

basin was important in controlling the position and shape of the Laramide trends, and an attempt has been made to contour it. The isopachs are merely suggestive and are superposed on the Algonkian isopachs of Figure 2. Information is very scarce in the western part of the area, but a measurement by Nolan (1935, p. 35) 60 miles west of the mapped area in the Gold Hill district is helpful. The northern Utah highland may have been out of water during most or all of Pennsylvanian time, and the northwest corner of the Colorado plateau subsided only slightly to receive a thin deposit of Pennsylvanian sediments (Baker, 1935, p. 1473-1475). Although 3000 to 6000 feet of Pennsylvanian sediments accumulated on the sites of the northern Wasatch and Uinta mountains, the major basin centered in the Oquirrh Range and possibly extended west–northwestward into Nevada. It crossed the neck of the deep Algonkian and Cambrian basin, producing there a combined late pre-Cambrian, Cambrian (Tintic quartzite), and Pennsylvanian thickness of about 30,000 feet. Gilluly lists 8900 feet of post-Tintic quartzite and pre-Pennsylvanian beds, making a total thickness of almost 40,000 feet of stratified rocks (excluding any possible Mesozoic) in the Oquirrh Range.

Near the eastern margin of the belt of deformation, south of the Uinta junction, thrust faults cut the thick Pennsylvanian series. The thrusts are associated with great overturned folds. The Nebo thrust is an example in the southern Wasatch, and Baker (oral communication) has recently noted thrust structures in the Provo-Spanish Fork section of the Wasatch. It appears that the Pennsylvanian strata yielded to compression characteristically by open regular folding except along the eastern margin of the basin and that the position and trend of the folds were influenced by the Colorado plateau, the northern Utah highland, and the Cottonwood and Sheeprock uplifts.

Mesozoic basins.—The present western limit of Mesozoic strata has been drafted on the isopach map (Fig. 2). Although the Cretaceous shoreline shifted through an east–west direction of about 50 miles (Spieker and Reeside, 1926) the present western limit of the Cretaceous rocks marks rather closely a line west of which a land of considerable relief terminated the areas of deposition of upper Cretaceous time (Spieker and Schoff, 1936, p. 104). The land area to the west stood high enough at times to furnish cobble and boulder conglomerates (Spieker and Reeside, 1926, Fig. 2; Schneider, 1929, p. 89), and immediately to the east the crust subsided to true geosynclinal depths (Spieker and Schoff, 1936, p. 104; Schoff, 1938, p. 378-379). In the Cedar Hills, between the southern Wasatch Mountains and the Wasatch Plateau, there is probably as much as 22,000 feet of Upper Cretaceous beds. No Mesozoic beds occur

west of the Wasatch in the area of this report, but the fact that the Triassic and Jurassic formations are not coarser than sandstone and that some are made up chiefly of limestone and shale suggests a shoreline somewhat farther west. Schuchert's paleogeographic maps (1923, p. 226, fig. 13) show that various seas invaded Utah from the west during Triassic time and that the site of the Wasatch Range was then not a dominant shoreline. The paleogeographic maps of the Jurassic, however, show western Utah to have been land and the east-facing shore to have been not far west of the present limit of Jurassic sediments. By Jurassic time, then, the western margin of the Colorado plateau and the Wasatch Range south of the Uinta junction was becoming defined as a structural lineament along which a prolonged and complicated orogenic record grew. About 10,000 feet of Upper Jurassic shale accumulated in the southern Wasatch Mountains and the Gunnison Plateau (Eardley, 1932, p. 331; Schoff, 1938, p. 377). Although undergoing some epeirogenic movements, the Colorado plateau acted as a buttress or foreland to the flanking geosynclinal sediments, which were apparently under pressure also from rigid highland masses to the west. Spieker and Schoff (1936, p. 104) have discovered evidence of an orogenic movement of post-Jurassic and pre-Upper Cretaceous age in central Utah and have dated the main orogeny between middle and late Montana time.

North of the Uinta junction in the Wasatch Range the present western margin of the Mesozoic beds is mostly covered by Tertiary sediments, and the orogenic history is not well known. However, several episodes of deformation in late Cretaceous and early Tertiary time occurred in the Coalville-Morgan area. Cretaceous sediments accumulated to thicknesses of 15,000 feet (Mansfield, 1927, p. 377) in southeastern Idaho and of more than 20,000 feet in southwestern Wyoming (Veatch, 1907, p. 59). Mansfield (1927, p. 377) notes that the Cretaceous beds in southeastern Idaho disclose at least two epochs of deformation and erosion.

The thick accumulations of the Idaho and Wyoming sediments in Cretaceous time indicate the proximity of rising land areas just before and during the Cretaceous. According to Mansfield (1927, p. 194), the Sierra Nevada movement (post-Jurassic) had elevated the uplands to the west so that they furnished large quantities of coarse debris to the streams, but the attitude of the Jurassic beds, which were covered by these gravels, was little changed by the disturbance. The material in the conglomerate shows that the mountains that supplied the debris were composed chiefly of Paleozoic sedimentary rocks but may have included some small areas of Triassic or Jurassic beds. The Cretaceous strata themselves were soon compressed and folded in the Laramide revolution, and before the Laramide orogeny was over some of the lower Tertiary beds were disturbed.

Idaho-Wyoming arc.—A map (Fig. 3) has been assembled to show in simplified form the structural features of the north-central Utah area of this report together with those of adjacent related areas in Idaho,

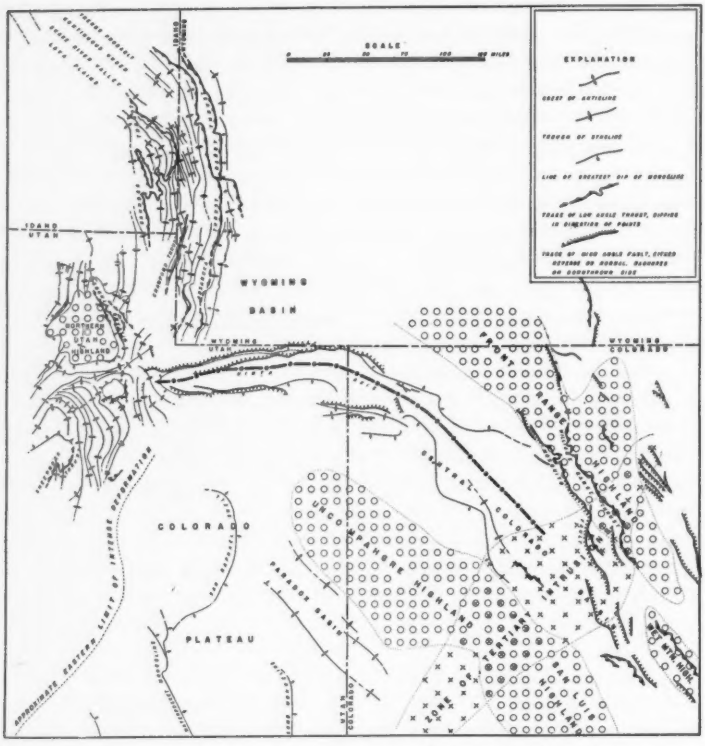


FIGURE 3.—*Map showing relation of the Laramide structural trends of north-central Utah to those of Idaho, Wyoming, and Colorado*

Wyoming, and Colorado. The folded and thrust structures of southeastern Idaho, western Wyoming, and northeastern and central Utah form a great arcuate belt known as the Idaho-Wyoming arc. The north end of the arc begins northwest of the Snake River lava plains at the center of the eastern margin of the Idaho batholith (King, 1933, Pl. 1). Southward the Idaho-Wyoming arc meets acutely the west end of the arcuate uplift of the Uinta Mountains. Just northwest of the junction lies the small fairly rigid mass of the northern Utah highland directly

in the path of the southwest-trending structures. The buttress deflects the main folds around it, and its resistance to deformation has resulted in the Willard thrusts across the folds on its northeast border. The northward-trending folds and thrusts of the Wasatch Mountains south of the junction swing westward and join the folds of the Idaho-Wyoming arc. Thence together they continue southwestward to the Sheeprock uplift and, perhaps, even to the western border of Utah south of the Gold Hill mining district. The Idaho-Wyoming arc is clearly one of the largest structural elements of the western Cordillera. It extends 400 miles from north to south and has a radius of curvature of about 225 miles. It was formed by horizontal pressure applied from the west (Mansfield, 1927, p. 386). The total amount of crustal shortening according to Mansfield (1927, p. 386) and Veatch (1907, Pl. 4) is about 35 miles, although King (1933, p. 35) gives "as much as 40 miles."

Wyoming basin and Colorado plateau.—If the Uinta uplift be momentarily ruled out of the picture (Fig. 3), the Wyoming basin [1] (Green River basin) and Colorado plateau would be seen to merge as a structural unit. The eastern limit of the Idaho-Wyoming arc appears to end rather abruptly against the slightly deformed sediments of the Wyoming basin (Veatch, Pl. IV). Butler (1929, p. 35), Spieker (1931, p. 52), and others note the abrupt margin of the comparatively flat-lying and thin sediments of the Colorado plateau against the highly folded sediments to the west. The eastern limit of deformation south of the Uintas, if projected northward across the Uinta uplift, meets the eastern limit of the Idaho-Wyoming arc. Butler (1929, p. 35) and King (1933, p. 31) believe that the crystalline complex of the Colorado plateau was covered by only a moderate thickness of strata, part of which was removed during successive periods of uplift and erosion. Although not strongly positive in character, the plateau acted as a fairly resistant buttress to deformation during the Laramide revolution.

During the Laramide revolution an east–west zone of weakness traversed the northern part of the buttress, and in this zone a fairly thick series of sediments was deformed into the block-shaped uplift of the Uinta Mountains. Forrester (1937, Fig. 1) believes that the area of the Uinta uplift was a geosyncline during Algonkian, Paleozoic, and

[1] The term "Wyoming basin" is a physiographic designation (Physiographic map of the U. S., by N. M. Fenneman and D. W. Johnson) for the elevated interior plains of Wyoming and has been chosen in preference to "Green River basin." The Green River and Uinta basins subsided chiefly following the Laramide revolution and hence were not features of the pre-Laramide setting. In formulating the Laramide controls the Tertiary basins are, therefore, not considered, and it would be misleading to speak of the rigid block north of the Uinta Mountains as the Green River basin when the basin was not yet in existence. The physiographic term "Wyoming basin" corresponds with the physiographic term "Colorado Plateau" and both may be used also in a structural sense.

Mesozoic time, that the sediments are much thicker there than to the south in the Colorado plateau and to the north in the Wyoming basin, and that the area was established as a zone of weakness in Algonkian time. These conclusions are based on his isopach map. The insufficiency of data leads Spieker (personal communication) to doubt the conclusions. It appears to the writer that the late pre-Cambrian strata thin markedly eastward from the Sheeprock uplift to the southern Wasatch Mountains (15,000 feet to 500 feet) and they may wedge out entirely eastward on the margin of the plateau. As there are no outcrops in the plateau area between the Uintas and the Grand Canyon the problem must remain largely speculative.

The relative amount of horizontal movement of the Colorado plateau toward the Wyoming basin has been approximated by the writer as 7 miles. This is but a fraction of the amount of compression that occurred in the sediments bordering the western margin of the nuclear areas. The fault along the northern flank of the Uintas has been called a thrust and, although definitely a very high-angle fault, has been misinterpreted by some to signify considerable horizontal contraction.

The history of the Colorado plateau has not been so uniform as the foregoing generalized statements suggest. The main pre-Laramide features of the plateau region and adjacent related areas were the Paradox basin, the Uncompahgre highland (Baker, 1935, p. 1488), the Central Colorado basin, and the Front Range and Wet Mountain highlands (Lovering, 1932, p. 20). (*See* Fig. 3.)

The elevation of the Uncompahgre highland, according to Baker (1935, p. 1495), began in late Mississippian or early Pennsylvanian time and continued until the close of the Carboniferous. Southwest of the highland the crust subsided and the salt-bearing sediments of the lower Pennsylvanian Paradox formation and the Hermosa formation accumulated. By upper Triassic time the highland had subsided sympathetically with the area to the west and was buried by the Chinle formation of Upper Triassic age. Subsidence of the whole area until the close of the Jurassic followed, with gentle westward tilting so that there was a slight uplift in western Colorado.

The Central Colorado basin, according to Lovering (1937, p. 20-21) and Burbank (1933, p. 279-283), was a trough during the Paleozoic era and received as much as 15,000 feet of sediment in its southern half. The accumulation of the thick sediments involved the downwarping of the trough and the uplift of the adjacent highlands. The Front Range highland was present as early as Cambrian time, remained high, and suffered erosion until the Dakota sea invaded the region. The Uncompahgre

highland did not appear, according to the sedimentary record in the central Colorado basin, until Devonian time.

The highlands and basins in Colorado trended northwest–southeast, and their positions profoundly influenced the Laramide structures (Bur-

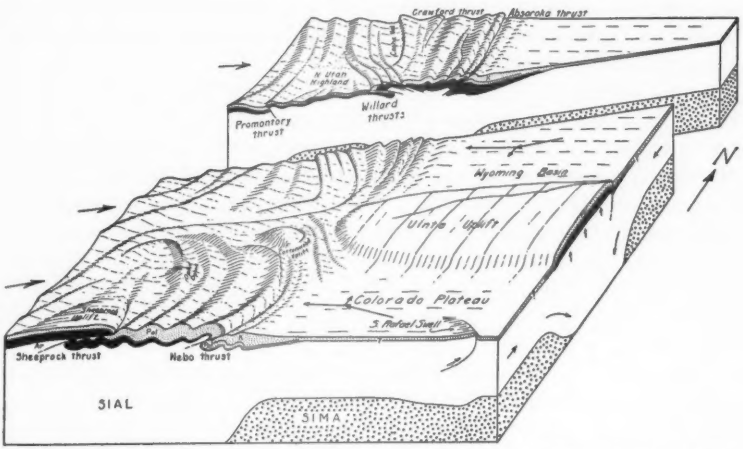


FIGURE 4.—*Block diagram of the Laramide structures in the Wasatch-Great Basin region*

bank, p. 283-288). During the Laramide orogeny intense deformation was localized at the edges or near the centers of the deepest basins of deposition in Colorado, with the formation of folds and of great thrust faults. These took the northwest course of the highlands. The Colorado plateau in Utah did not stand entirely unyielding to the Laramide forces that deformed the crust so intensely on either side, for within it there appeared broad asymmetrical folds, such as the San Rafael Swell, the Circle Cliffs or Water Pocket monocline, and the Kaibab and Monument upwarps (Baker, 1935, p. 1498). The uniform character of the folds, with steeply dipping east flanks, gently dipping west flanks, northerly trend, and with ends that flatten as they veer westward, has led Baker (1935, p. 1502) to believe that the structures are due to east–west compression. The more competent basement rocks transmitted the Laramide forces for long distances but broke along a few planes of weakness to form thrust faults that have the aspect of ramps. The faults are almost vertical near the surface, but Baker postulates that they dip westward at depth and become more gently inclined. The strata near the

surface were more pliable and flexed over the upthrown crystalline blocks. (For example of this kind of structure, *see* Thom, 1937, p. 1-7.)

The smaller folds in the Paradox basin, although formed first in early Mesozoic time, were either accentuated by the Laramide forces or influenced the localization of other small folds nearby. The beds overlying the Paradox formation were supported by relatively plastic rock salt and hence buckled easily against the buttress of the crystalline rocks of the Uncompahgre highland.

As a summary of the Laramide structures of the Wasatch-Great Basin region, a block diagram is presented (Fig. 4). It assumes no erosion and no Tertiary beds. It depicts the crust hypothetically to a depth of about 30 miles in order to show Lawson's (1931, p. 264-267) conception of the isostatic conditions of the Uinta Mountains and also to indicate the thickness of the sedimentary rocks relative to the granitic (sial) and basaltic or peridotitic (sima) shells. The interpretation of all other structures at depth and the suggested resolution of forces are the writer's.

BASIN AND RANGE TRENDS

DEFINITION

The Basin and Range fault system is clearly younger than the Laramide folds and thrusts. Its individual faults are generally and most easily recognized by topographic unconformities between the ranges and the alluviated valleys. Schneider (1925, Fig. 2), Gilluly (1932, p. 81-88), Marsell (1932, p. 90-93), and the writer (1934, p. 388) have confirmed Gilbert's (1928, p. 1-67) and Davis' (1903, p. 129-174; 1905, p. 17-56) conclusion of block faulting from physiographic evidence by finding actual exposures and by measuring stratigraphic displacements. Faults thus established were transposed directly to the regional map (Pl. 1). The interpretation of the Wasatch fault on Plate 1 is that of the writer and is slightly at variance with Gilbert's (1928, fig. 12), particularly at the Salt Lake salient (T. 1 N., R. 1 E.) and in the southern Wasatch Mountains. The reconnaissance geologic map of Utah prepared by Butler and others (1920, Pl. IV) shows Basin and Range faults along the west face of the Oquirrh and Stansbury ranges. Gilluly's subsequent detailed interpretation of the faults along the south half of the Oquirrhs has been transcribed. The north half is the writer's interpretation, as is also the detail along the west front of the Stansbury Range.

The map of the region (Pl. 1) is admittedly incomplete, as many Basin and Range faults not active in recent geologic time are masked by valley fill. Gilluly and the writer describe faults running out into the alluvium where they are lost. An earthquake occurred north of Great Salt Lake in Hansel Valley (T.7 and 8W., R.12 and 13N.), at which time scarps

were formed in the alluvium almost in the center of the valley. A complete map of the Basin and Range faults will not be available until the region has been prospected seismologically to locate buried fault scarps.

DESCRIPTION

The dominant feature of the Basin and Range system is the Wasatch fault (Gilbert, 1928, Fig. 12). It is believed to be continuous from T. 8 S. to T. 11 N., a distance of 115 miles, except for an interruption at the Salt Lake salient (T. 1 N.). Its trace is slightly sinuous, but the small irregularities fall within a fairly straight course that runs about 12 degrees west of north. The writer believes that the Wasatch fault dies out north of Emigration Canyon and begins again about 6 miles to the north. Across the Salt Lake salient there is no known structural or physiographic evidence of the fault (Schneider, oral communication), although Gilbert there maps the fault as unbroken. An offset fault farther west that bridges this gap extends through Warm Springs and Beck's Hot Springs (Pack, 1926, p. 399-410) northward to St. Joseph, a distance of about 6 miles. (*See* Plate 1.) The relative simplicity of the Wasatch fault is disrupted in the southern Wasatch Mountains (Eardley, 1933a, map 11 and fig. 10) where a group of faults has an almost en echelon pattern.

The writer has found the maximum throw of the fault in the southern Wasatch to range from 5000 to 6000 feet and to be about 3500 feet in the Farmington section. Marsell (1932, p. 107-109) believes that the maximum throw in the central Wasatch is 5000 to 6000 feet. The dip of the fault plane ranges from 50 to 72 degrees. Schneider (1925, Fig. 2) pictures a slickensided surface exposed by removal of Lake Bonneville sediments that has a dip of 72 degrees. The fault plane dips 50 degrees in two places in the southern Wasatch Mountains (Eardley, 1933a, p. 388-389). Gilbert and Davis believed that the low-dipping triangular facets of the eroded fault scarp indicate the dip of the fault plane. The problem has been reviewed by Gilluly (1928, p. 1103-1130) for the Oquirrh Range, and he concludes that the scarps dip at lower angles than the faults. The evidence from a few actual outcrops along the Wasatch front sustains this conclusion.

In the area of the Pennsylvanian basin a fairly distinct en echelon arrangement of Basin and Range faults is evident. (*See* Plate 1.) This includes the Basin and Range faults of the Oquirrh and Stansbury ranges and the faults of the southern Wasatch Mountains. The zone trends N. 20° W. in the southern Oquirrhs but runs more nearly north along the range as a whole. The individual faults composing the zone trend N. 20°–50° W. A northward trend is present in the Stansbury Range.

The displacements are distributed among the step faults of the en echelon plan and attain a maximum of about 3000 feet. The faults die out at both ends and have lengths of 3 to 12 miles in the Oquirrhs and are possibly longer in the Stansbury Range. Gilluly (1932, p. 85) measured the dip of faults at 14 places in the southern Oquirrhs and found it to range from 40 to 64 degrees and to average 57 degrees. The dip of the West Mercur fault changes from 40 to 60 degrees within a mile. Other faults have comparable variations showing that the surfaces veer in vertical as well as in horizontal direction. Some of these faults that are reflected in the present topography may be rejuvenated older faults (Gilluly, 1932, p. 84).

East of Morgan Valley the foot of the mountains is marked by a fresh scarp. Schneider (1925, p. 41-43) and Bailey (1927, p. 5-8) describe Basin and Range faults in Cash Valley, but a recent field conference of geologists of the Utah Academy of Science failed to find convincing evidence.

Promontory Range does not display young fault scarp topography, but numerous warm mineral springs along the southeast base suggest faulting. If faulting has occurred, it is of such antiquity that the recognizable features of a fault scarp are lost. The same is true of Antelope Island and Stansbury Island.

The Basin and Range faulting is post-Laramide, post-Wasatch conglomerate, post-mineralization, and is younger also than an extensive early Tertiary erosional interval. In several places within the area of this report faulting continued after Lake Bonneville time, and earthquakes are believed to betray its recent activity. On the basis of physiographic evidence in the Oquirrh Range, Gilluly thinks that faulting began not later than middle Pliocene (1932, p. 40, 77, 86). The writer believes that in the Wasatch it began in late Pliocene and proceeded through the Pleistocene to the present (1933a, p. 394-397).

RELATION TO LARAMIDE AND PRE-CAMBRIAN TRENDS

The regional deforming forces of the Laramide revolution probably acted in an east–west direction, and the general trend of the Basin and Range faults is north–south. From this it has been inferred that the two structure systems are closely related in origin (Hobbs, p. 134). The most striking fact that comes from a study of the map (Pl. 1), however, is the lack of alignment of the Basin and Range and Laramide structures. The faults cut the folds from acute to right angles and in other places are parallel with them. Without meaningful deflections, they cut through the Cottonwood uplift with its granitic intrusions and thick quartzite series. They cut the hard Farmington Canyon complex with approxi-

mately the same displacement as where they cut the sedimentary veneer. They transect the Willard thrust slices obliquely.

It appears, nevertheless, that the Laramide structures locally influenced the course of the later Basin and Range faults. In the southern Wasatch Mountains the major Laramide structure is an asymmetrical and overturned anticline. The Wasatch fault follows the axial plane of the fold for several miles. The fault also parallels the fold in the Mt. Timpanogos section of the Wasatch Mountains. The belt of quasi en echelon faults of the southern part of the Oquirrh Mountains follows approximately the irregular trend of the folds. However, in view of the fact that these faults follow the Laramie folds for only short distances and elsewhere show a random cross-cutting pattern, it is believed that a Laramide control is at best incidental and local.

Some of those who have related the Basin and Range fault system to the Laramide structures may have failed to realize that in the Utah area the Laramide deformation probably ended before the close of Eocene time. Yet, the Basin and Range orogeny began not earlier than the Pliocene. At least half the Tertiary period intervened between the two.

In relating the Basin and Range and the Laramide structural trends to the pre-Cambrian trends, two pre-Cambrian orogenies must be distinguished. The folds and foliation of the old crystalline complex were formed chiefly before the late pre-Cambrian sands, silts, and clays were deposited (Eardley and Hatch, 1939). These sediments were then deformed (at least around the northern Utah highland) before the Lower(?) and Middle Cambrian sands were deposited. These are the most evident diastrophic events of the pre-Cambrian history.

The foliation of the metamorphosed sediments of the crystalline complex is generally parallel with their bedding, and therefore the strike of the foliation is approximately the direction of the fold axes. Several folds have already been described east of Bountiful and Farmington whose axes trend N. 60° W. This strike continues northward across Weber Canyon. Still farther north in Ogden Canyon schlieren in a grey granite strike due north. In Willard Canyon at the north end of the highland the strike of the foliation is N. 23° W. On Antelope Island the foliation runs generally due north, but at the south end it strikes east of north. This is the only location where the foliation has been observed to strike east of north. It has been recorded as N. 4° W. in the southern Wasatch Mountains (Eardley, 1932, p. 311).

Two trends in the crystalline rocks seem dominant, one due north and the other N. 60° W. Since large areas of the old highland are covered with Pleistocene and Recent alluvium it would not be safe to conclude that these are individualistic basic trends, especially in view of the fact

that there are many variations ranging from one trend to the other. The Basin and Range faults cut the crystalline foliation at angles of a few degrees to 60 degrees. The greatest divergence occurs in the Bountiful-Ogden area, the least divergence at the south end in the southern Wasatch Mountains and at the north end in the northern Wasatch. The meaning of these details is not known, but it appears that the structures in the old crystalline rocks inherited from early or middle pre-Cambrian time exerted little influence on the course of the Basin and Range faults. Both structure systems, however, trend in general west of north—the crystalline foliation as much as N. 60° W., and the faults N. 15° W.

The Laramide trends follow the crystalline rock formation with only small angular discordance in the southern and northern Wasatch Mountains and for a distance along the eastern edge of the Farmington highland. A Laramide fold axis on Antelope Island is approximately parallel with the foliation of the crystalline rocks. At all other places the two trends diverge widely, and the earlier structure appears to have exerted no control over the later. However, the northern Utah highland, a fairly resistant buttress, profoundly influenced the course of the Laramide structures, although the ancient foliation within it did not control, except perhaps locally, the actual trend and position of the folds and thrusts. It does not appear, on the other hand, that the Farmington highland, even as a mass, had any influence on the course of the Basin and Range faults.

A theory of the sedimentary, deformational, and erosional history of late pre-Cambrian and Cambrian time has already been outlined in this paper. The writer postulates that the highland had considerable relief during the early part of late pre-Cambrian sedimentation but before the close of the pre-Cambrian was buried by about 3000 feet of strata. At the same time the area to the southeast and east subsided, permitting sediments to accumulate into a thickness of more than 11,000 feet. Uplift of the highland then occurred, and the 3000 feet of strata over the highland were stripped off by erosion, except for small patches. The structure thus produced appears to be simply a great upwarp of the late pre-Cambrian beds around the highland. The Laramide structures follow closely the margin of the late pre-Cambrian upwarp, except some of the westward deflections of the Willard thrust sheets, and, therefore, it appears that the eastern margin of the Farmington highland was established first in late pre-Cambrian time and that the marginal structures were accentuated during the Laramide revolution. Undoubtedly structures in the late pre-Cambrian sediments other than the upwarp around the highland were formed in pre-Cambrian time.

A suggestion of this is seen in the unconformity in the southern Wasatch Mountains where the Cambrian quartzite rests on 500 to 1000 feet of late pre-Cambrian sediments. The locality is south of the intervening Pennsylvanian and late pre-Cambrian basins and may represent a marginal upwarp and overlap on the positive area of the Colorado plateau. Folds of pre-Cambrian age in the late pre-Cambrian sediments also may occur, but the few scattered exposures leave the suggestion without proof. The influence of the early folds, if they exist, on the Laramide trends is not known.

As far as can be seen the late pre-Cambrian trends exerted little influence upon the course of the Basin and Range faults.

CAUSE OF BASIN AND RANGE FAULTING

Two explanations for the Basin and Range faults were ventured by Gilbert and Le Conte in 1875 and 1889 respectively, but the solution of such a major problem has since generally been regarded as one that must await detailed information of the province as a whole together with a synthesis of the geology of the entire western Cordillera. The hypotheses presented to date are briefly as follows:

Gilbert (1875, p. 60) supposed that the region was broken into a series of blocks by vertical pressure from below. The vertical or nearly vertical adjustments of the brittle surface rocks were due to folds in a deep-seated layer that were induced by regional compression.

Le Conte (1889, p. 257-263) suggested that the region was first raised by volcanic forces. When the upward pressure was relieved by the extrusion of vast amounts of lava, the weight of which was added to the overlying portion, subsidence of the whole region occurred. The subsidence was in the nature of a collapse in which the crust broke into blocks that settled irregularly.

Ransome (1915, p. 243) agrees with Le Conte that the structures were produced by a collapse of the region rather than an elevation but sees no reason for supposing that any elevation by volcanic forces preceded the collapse. Butler (1920, p. 105) also believes that settling of the region is the most rational explanation. He cites a supposedly analogous condition of a pond in which the water has drained from beneath an ice sheet.

> "In settling the ice sheet forms monoclinal folds which pass into faults, faults of small displacement come together in faults of greater displacement, and where settling is pronounced a structure comparable to that of the Basin Ranges may develop."

Gilluly (1932, p. 86-87) investigated the isostatic effect of erosional transfer of load in the Oquirrh Mountains which, according to the theory of isostasy, results in local "unloading" and "loading." Some geologists

had suggested (apparently orally) that erosion from the mountain blocks and deposition upon the valley blocks may have led to the rising of the mountains and sinking of the valleys. As Gilluly points out, such a process offers no key to the cause of the initial faulting, but there can be no doubt that the tendency of the transfer of load is to continue the movement. He concludes that the tendency toward isostatic adjustment between mountain range and adjacent valley had little effect in producing the observed displacement between them. Gilbert (1890, p. 365-392) also thought that individual Basin Ranges are sustained by the strength of the earth's crust and that they are not individually balanced. He arrived at this conclusion from a study of the loading effect of the water of Lake Bonneville and the lack of response to this loading through faulting, such as would be expected on an assumption that there was isostatic control of individual Basin Ranges.

CONCLUSIONS

The uplifts—the Uinta, Cottonwood, and Sheeprock—form an east-southwest arcuate trend and one of the major structural elements of the region. The Uinta is a long fairly flat-topped arch with steeply dipping and faulted flanks; the Cottonwood is a dome; and the Sheeprock a complexly folded and faulted portion of the crust. In the core of each a thick series of late pre-Cambrian beds are exposed. The Cottonwood uplift is separated from the Uinta by a small synclinorium with many high- and low-angle faults. The Cottonwood uplift is separated from the Sheeprock by another synclinorium. There is no fold connecting one uplift with another across the synclinoria. The Park City and Cottonwood intrusions line up along the westward projection of the Uinta axis, but those of the Oquirrh Range lie on either side.

The northern Utah highland is a small area of general positive character just northwest of the Wasatch-Uinta junction. It has also acted as a fairly rigid buttress, and between it and the Cottonwood uplift a thick sequence of beds has been rather closely folded. The buttress lies directly in the path of the folds and thrusts of the Idaho-Wyoming arc and strikingly deflects the folds around it. Along its northeast margin the Willard thrusts have formed athwart the folds of the arc. Although the northern Utah highland deflected the folds of the Laramide revolution around it, the internal pre-Cambrian structures of the buttress only appear to have controlled the Laramide structures locally.

The highland is believed to have been buried by about 3000 feet of late pre-Cambrian sediments while to the east, southeast, and south subsidence and accumulation of 10,000 to 15,000 feet of sediments occurred. The highland was then elevated and the 3000 feet of strata was mostly stripped

off by erosion. Subsidence followed with the deposition of Lower and Middle Cambrian sands in overlapping fashion on the upturned and truncated edges of the late pre-Cambrian strata around the margin of the highland and on the ancient crystalline rocks of the core of the highland. The dips of the upturned pre-Cambrian beds around the highland were further increased by uplift in Pennsylvanian, Mesozoic, and Laramide times.

A basin in which about 20,000 feet of strata accumulated formed during the Pennsylvanian period south of the northern Utah highland and north of the Sheeprock uplift. Its axis trends west–northwest almost at right angles to the axis of the late pre-Cambrian and Cambrian basin in this area. The late pre-Cambrian and Paleozoic sedimentary veneer was probably 40,000 feet thick in the center of the basin. The characteristic structures of the basin are long regular folds, generally open, but some overturned and closed. Near the eastern margin of the Pennsylvania basin where recently a thick Permian section has been discovered great overturned folds and thrust faults displace the beds.

By Jurassic time the western margin of the Colorado plateau and the Wasatch Range south of the Uinta junction was becoming defined as a structural lineament along which a prolonged and complicated orogenic record grew. The site of maximum subsidence shifted eastward from the trough of the Pennsylvanian basin, and about 10,000 feet of Jurassic sediments was deposited in the southern Wasatch Mountains and Gunnison plateau area. Just east of the southern Wasatch as much as 22,000 feet of Cretaceous sediments accumulated. Although undergoing some epeirogenic and orogenic movements, the Colorado plateau acted as a buttress or foreland to the flanking geosynclinal sediments which suffered compression during the Laramide revolution from rigid highland masses to the west.

North of the Uinta junction great thickness of Cretaceous sediments also accumulated. The Sierra Nevada movement had elevated uplands to the west, and the Cretaceous strata were soon profoundly disturbed. The Wyoming basin acted as a foreland buttress. The orogeny continued into the Tertiary period.

A long episode of epeirogenic adjustments and erosion followed the Laramide revolution. In Pliocene time the crust from the Wasatch Mountains westward was broken by the faults of the Basin and Range orogeny. The dominant characters of this structure system in the Wasatch-Great Basin region are as follows:

(1) A master fault—the Wasatch fault—115 miles long, with displacement of 3000 to 6000 feet, forms the eastern limit of the faulted part of

the area. The Morgan Valley fault is exceptional—all other definitely established faults lie to the west.[2] The Wasatch fault trends about 12 degrees west of north. In places its displacement is probably distributed among step faults. It is a dip-slip normal fault and dips 50° to 70° W.

(2) A quasi en echelon pattern of smaller normal faults spreads over the deep basin of Pennsylvanian sediments. The pattern may extend north and south of the basin area, but evidence for this is lacking. The en echelon zones trend a little west of north, in about the same direction as the Wasatch fault, but the individual faults strike N. 20°-50° W. The displacements are commonly distributed in step faults which dip on an average of 57° W. The fault surfaces are usually somewhat sinuous in both vertical and horizontal directions.

(3) The major blocks included between the faults all seem to have been tilted 2° to 4° eastward. One small questionable fault (T.7S., R.6W.) has downthrow on the opposite side.

(4) The Basin and Range faults are not aligned with the pre-Cambrian or Laramide structure systems. The only control that the older structural trends exerted on the course of the younger faults was very local. Neither have the northern Utah highland, the intrusions of the Cottonwood uplift, nor the late pre-Cambrian basins influenced perceptibly the course or the throw of the faults. The thick series of Pennsylvanian rocks may have been influential because the faults have an en echelon arrangement in the area of the Pennsylvanian basin. The pattern is not conspicuous along the Wasatch fault outside of the area of thick Pennsylvania sediments.

(5) The width of the tilted fault blocks ranges from 4 to 24 miles, but a fairly uniform width of 18 to 24 miles is found in the four major blocks of the area—the Wasatch, Oquirrh, Stansbury, and Cedar[3] mountains blocks.

(6) The faults of the High Plateaus of Utah, just south of the area of this report are associated with monoclinal flexures but neither Butler (1920, p. 105) nor the writer (1934, p. 157-158) believe that they differ otherwise from Basin and Range faults. The folded and less pliable Paleozoic and late pre-Cambrian limestones, sandstones, and quartzites, and in places the earlier pre-Cambrian crystalline rocks of the Basin and Range province, are generally faulted, but the younger Mesozoic and Tertiary sediments, in which the High Plateaus contain much shale, are commonly flexed into monoclines. The same orogenic stresses that caused

[2] It is not known definitely whether the Wasatch fault continues southward into the High Plateaus area, but there large faults exist east of the longitude of the Wasatch fault.

[3] The Cedar Mountains are immediately west of the Stansbury Range and off the map (Pl. 1). A large fault is shown along their western base on Plate 4 of U. S. Geological Survey Professional Paper 111. The Wasatch block is postulated to be about 18 to 20 miles wide (Eardley, 1933, p. 265).

the faults of the eastern part of the Basin and Range province probably caused the faults and flexures of the High Plateaus.

(7) If it is proposed, however, to include the High Plateau faults in the Basin and Range group, a conspicuous exception to one of the characteristics of the Basin and Range faults of the Wasatch Great Basin region may be noted. At least three major faults in the High Plateaus have downthrow on the east, and the related fault blocks probably were tilted toward the west. Gilbert (1928, p. 70-74) records a fault of this type along the east base of the Fish Spring Mountains (T. 12 S., R. 14 W.), and other unknown or buried fault scarps in the eastern Great Basin may have downthrow to the east.

(8) Since the faults occur in extensive areas where no lavas have been erupted, it seems improbable that volcanism, except locally, caused the collapse of the surficial crustal rocks.

(9) Certain students of Basin and Range structure have conceived of horst and graben faulting as characteristic. Gilbert refers to the Wasatch Mountains as a horst (1928, p. 62), and Louderback (1923, fig. 1) reproduces an idealistic figure of Keyes showing a horst and a graben. Gilbert's horst of the Wasatch has been contested (Eardley, 1933b, p. 257-265), and the writer is not certain of any examples of graben or horst structure west of the Wasatch in the eastern part of the Great Basin. He is strongly inclined to believe that almost all the fault blocks were rotated, downward on one side and upward on the other. Marysvale and San Pete valleys in the High Plateaus may be asymmetrical grabens, but the downthrown blocks probably tilted as they sank.

The lack of influence of surficial rock masses and their pre-existing internal structures on the Basin and Range faults leads the writer to the conclusion that the local causal forces were extremely deep-seated. The movements probably were not due to a regional compression that "folded" the deep-seated material because the more brittle surficial rocks could not easily have escaped some compression even though deformation of the deep-seated rocks resolved the horizontal forces in part into vertical forces. There is no compelling evidence of compression connected with any of the Basin and Range faults familiar to the writer. In fact, it is difficult to escape the idea that they mean an extension of the crust.

Cloos (1929, p. 233-241, Pls. 7-13) has experimented with models of thin clay and by horizontal tension has produced structures that in certain aspects are similar to the Basin and Range fault blocks. Additional experimental work with scale-models, according to the recent paper of Hubbert (1937, p. 1459-1519), appears to offer an interesting and fruitful approach to the problem of origin. The writer feels that an experimental

imitation of subcrustal flow to determine its effects on a sial that is covered with a variable thickness of sediments would lead to worthwhile results.

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Manuscript received by the Secretary of the Society, January 6, 1939.
Presented before the Geological Society, December 28, 1938.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1311-1322 AUGUST 1, 1939

TWO LOWER PALEOZOIC GROUPS, ARBUCKLE AND WICHITA MOUNTAINS, OKLAHOMA

BY CHARLES E. DECKER

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ABSTRACT

This paper establishes two groups—Timbered Hills and Arbuckle—for the lower Paleozoic formations of the Arbuckle and Wichita mountains of Oklahoma. The Timbered Hills group is divided into three formations, and the Arbuckle into nine formations and two members. A brief description is given, and a type section designated for each formation. The three formations of the Timbered Hills group and four in the Arbuckle group are assigned to the Upper Cambrian, and the other five formations of the Arbuckle group to the Lower Ordovician. A table shows the classification proposed by the writer compared with earlier classifications.

INTRODUCTION AND ACKNOWLEDGMENTS

It seems desirable to make as systematic a classification of the formations of these groups as is possible at present. Doubtless, important additional information on the ceratopeas of the Arbuckle limestone will be contributed by Josiah Bridge, for whom the writer supplied a large number of forms collected from specific zones in many sections. It is

hoped that some particularly good correlative evidence will be secured by Edward Frederickson, who with the writer made extensive collections of trilobites from numerous zones in the Honey Creek formation and in the lower part of the Arbuckle limestone.

The writer will present two groups of Lower Paleozoic rocks of the Arbuckle and Wichita mountains and will show the divisions of these groups into formations, with brief descriptions and the assignment of a type section for each formation. Formation names and type sections by Ulrich and Taff will be used insofar as possible, but several of the names used by Ulrich (Ulrich and Cooper, 1938, p. 23) are not desirable because of previous use, as indicated by Wilmarth (1938) in the new lexicon of geologic names. Of the two groups, the Timbered Hills and the Arbuckle, the former is used because both the Reagan and Honey Creek formations are well represented in and adjacent to the East and West Timbered Hills of the Arbuckle Mountains, and the Arbuckle group is used to include a series of nine formations into which the very thick Arbuckle limestone has been divided. In Table 1, the formations are placed in two divisions according to age. Three formations of the Timbered Hills group—Reagan, Cap Mountain, and Honey Creek—and four in the lower part of the Arbuckle group—Fort Sill, Royer, Signal Mountain, and Butterly—are assigned to the Upper Cambrian; and five formations of the Arbuckle group—McKenzie Hill, Strange, Cool Creek, Kindblade, and West Spring Creek—are assigned to the Lower Ordovician.

Of the 12 formations and 2 members shown in Table 1, No. 8, three in the lower part—Reagan, Cap Mountain, and Honey Creek—had been associated by the writer as in No. 5. The lowest of these—the Reagan—was placed by Taff in the Middle Cambrian, and he recognized an upper calcareous phase. Ulrich, as in Nos. 2 and 3, separated the upper calcareous phase as the Honey Creek member and placed both it and the Reagan in the upper instead of the middle Cambrian. As indicated in No. 4, he raised the Honey Creek from a member to a formation. Still later, Bridge and Ulrich shifted the Honey Creek into the basal part of the Arbuckle as shown in Nos. 6 and 7.

Six of the formations in the Arbuckle group in No. 8—Fort Sill, Royer, Signal Mountain, McKenzie Hill, Cool Creek, and West Spring Creek—had been associated in that group before as in No. 5 and had been listed by Ulrich (Ulrich and Cooper, 1938) as in No. 7. Also, four of them had been used by Bridge as in No. 6, and three of them—Fort Sill, Royer, and Signal Mountain—had been described by Ulrich as noted in No. 4.

As to the other three formations used in No. 5 and also listed in No. 7 —Chapman ranch, Wolf Creek, and Alden—it seems apparent that Ulrich intended to use Chapman ranch for the heavy beds of limestone above the upper dolomite rather than for the dolomite. Accordingly, Butterly is the name proposed for the upper dolomite. Chapman ranch is here retained for the limestones above the upper dolomite and is here designated the lower member of the McKenzie Hill formation. For the upper member of that formation McMichel is here proposed. Two others of Nos. 5 and 7—Wolf Creek and Alden—are preoccupied (Wilmarth, 1938, p. 28, 29, 2360, and 2361), so Strange is substituted for Wolf Creek, and Kindblade for Alden. Gasconade of No. 7 is not used in this paper because Bridge (1936, p. 983) indicated that its fauna is equivalent to that of McKenzie Hill.

Table 1 shows as well as possible in a brief table the varying assignments of age for the various parts of the Arbuckle limestone. Ulrich (Ulrich and Cooper, 1938, p. 23) uses Upper Cambrian, Ozarkian, Lower, Middle, and Upper Canadian for different groups of formations. Bridge (1936, p. 983) places four of the Arbuckle units in the Cambrian, two of the middle units and all of the Upper Arbuckle in the Ordovician. The classification proposed in this paper, in No. 8 of Table 1, aside from the Timbered Hills group, places four formations of the Arbuckle group in the Cambrian and five in the Ordovician.

Grateful acknowledgment is made to Robert Dott for suggestions in regard to the manuscript and table, and to Hugh D. Miser and Josiah Bridge for helpful criticism.

DESCRIPTION OF FORMATIONS

GENERAL STATEMENT

A brief description of each of the 12 formations is given, in succession from oldest to youngest, together with the range in thickness and the assignment of upper and lower limits as closely as possible, and a type section is designated for each formation.

TIMBERED HILLS GROUP

General statement.—As noted and shown in Table 1, this group includes the Reagan sandstone below, Cap Mountain in the middle, and the Honey Creek formation above.

Reagan sandstone.—The Reagan sandstone was named for the old town of Reagan, northeast of which a small outcrop of the formation occurs in the Tishomingo quadrangle, about 9 miles north and slightly

west of Tishomingo. However, the type section is taken 7 miles west of Reagan and 5 miles south and west of the village of Mill Creek, where it overlies the edge of the large Tishomingo granite outcrop. Taff's (1904, p. 31) brief description of the type section follows:

"Section of the Reagan Sandstone"

	Feet
"Quartzite and arkose conglomerate, composed of poorly sorted granitic materials	30
Coarse grit and sand, with some clay and green sand in upper part, generally well stratified	370
Thin bedded and laminated sandstone becoming calcareous in the upper part	60"

This gives a total thickness of 460 feet. The Reagan rests unconformably on the coarsely crystalline Tishomingo granite, and the upper limit is at the base of a sandy dolomite called by Ulrich (1932, p. 742) Cap Mountain. Other good outcrops of the Reagan occur not only adjacent to the two Timbered Hills in the Arbuckle Mountains, but also to the east, south, and far to the northwest of Blue Creek Canyon in the Wichita Mountains.

Cap Mountain formation.—This formation has very limited distribution in Oklahoma, occurring only near the east end of the Arbuckle Mountains, south of the village of Mill Creek. It is a sandy dolomite, gray to greenish pink, and 126 feet thick. It lies between the Reagan sandstone below and the Fort Sill above. Ulrich considers it a local extension of the formation by the same name in Texas.

Honey Creek formation.—This formation gets its name from Honey Creek, which flows across its outcrop nearly at right angles to the strike on the west edge of the East Timbered Hills in the southwest corner of sec. 36, T. 1 S., R. 1 E., about 8 miles southwest of Davis. It rests upon the coarse arkosic sandstone at the top of the Reagan and extends up to the dense limestones at the base of the Fort Sill formation.

The Honey Creek was raised from a member to a formation by Ulrich (1932, p. 742-743), and a description of the formation is given in two parts which total a thickness of 124 feet. The Honey Creek has for a time been included in the lower part of the Arbuckle limestone by the United States Geological Survey, but, as Ulrich (1932, p. 742) raised it to the position of a separate formation, it has been separated from the Arbuckle limestone, according to Wilmarth (1938, p. 1781).

Recently, Frederickson and the writer measured the type section on Honey Creek in greater detail and divided it into six zones (dip 36°,

1. Taff 1903		2. Ulrich 1911			3. Ulrich 1927		4. Ulrich 1932
Middle Cambrian	Upper Cambrian 700 feet	Ordovician					No table
Reagan sandstone calcareous above	Arbuckle limestone	Cambrian (?)	Ozarkian (?)	Canadian	Ozarkian Lower	Canadian Upper (wanting)	Lower (?), Middle, Upper	Arbuckle limestone	
		Upper Cambrian	Basal Arbuckle	Main part Arbuckle					
		Reagan sandstone	Honey Creek limestone member				Honey Creek limestone member		Raised Honey Creek to a formation, gave sections and described Fort Sill, Signal Mountain, and Royer.
			Reagan sandstone				Reagan sandstone		

* References to columns 1 to 9 are given under appropriate date in end-bibliography

| | | |

TABLE 1.—*Lower Paleozoic formations, Arbuckle and Wichita mountains, Oklahoma*

Earlier and proposed classification *

3. Ulrich 1927	4. Ulrich 1932	5. Decker 1933			6. Bridge 1936				7. Ulrich (Ulrich and Coop			
Arbuckle limestone	No table	Upper Cambrian ?	Timbered Hills group	West Spring Creek	Arbuckle limestone	Ordovician	Zones		Arbuckle limestone	Canadian	U	West S
				Alden								Alden
				Cool Creek			1–180	Undifferentiated Arbuckle limestone			M	Cool C
											L	Unnam
	Raised Honey Creek to a formation, gave sections and described Fort Sill, Signal Mountain, and Royer.	Ozarkian (?)	Arbuckle group	Wolf Creek (dolomite)						Ozarkian		Gascon Wolf C
				McKenzie Hill			181 ..... 182 183	Graptolites McKenzie Hill 1014 feet ..... ?				McKe
							184	Chapman Ranch 144 feet				Chapn
				Chapman Ranch (dolomite)		Cambrian	185	Unnamed unit		Upper Cambrian		
				Signal Mountain			206 215 217	..... ? ..... Signal Mountain 104 feet				Signal
				Royer (dolomite)			218 223	Royer 681 feet				Royer
				Fort Sill				Fort Sill 150 feet				Fort S
Honey Creek limestone member		Upper Cambrian ?	Timbered Hills group	Honey Creek				Honey Creek				Honey
Reagan sandstone				Cap Mountain								
				Reagan	Reagan sandstone 430 feet							

* ropriate date in end-bibliography.

TABLE 1.—*Lower Paleozoic formations, Arbuckle and Wichita mountains, Oklahoma*

Earlier and proposed classification *

	5. Decker 1933			6. Bridge 1936					(U
						Zones			
	Upper Cambrian ?	Timbered Hills group	Honey Creek	Arbuckle limestone	Cambrian	218 223	Royer 681 feet Fort Sill 150 feet Honey Creek	Arbuckle limestone	
			Cap Mountain						
			Reagan				Reagan sandstone 430 feet		
	Ozarkian (?)	Arbuckle group	Royer (dolomite) Fort Sill						
			Chapman Ranch (dolomite) Signal Mountain			185 206 215 217	Unnamed unit .....?..... Signal Mountain 104 feet		
	Lower (?), Middle (?), Upper Canadian		McKenzie Hill		Ordovician	181 ..... 182 183 184	Graptolites McKenzie Hill 1014 feet .....?..... Chapman Ranch 144 feet		
			Wolf Creek (dolomite)						
			Cool Creek			1–180	Undifferentiated Arbuckle limestone		
			Alden						
			West Spring Creek						

phy.

7. Ulrich (Ulrich and Cooper) 1938				8. Decker 1939			9. Decker 1928
							Zones
Arbuckle limestone	Canadian	U	West Spring Creek	Lower Ordovician	Arbuckle group	West Spring Creek	1–134
			Alden			Kindblade	135–153
		M	Cool Creek			Cool Creek	154–181
		L	Unnamed unit				
	Ozarkian		Gasconade Wolf Creek			Strange (dolomite) (in Wichitas only)	
			McKenzie Hill			McKenzie Hill	
						a. McMichel	182–183
			Chapman Ranch			b. Chapman Ranch	184
	Upper Cambrian			Upper Cambrian		Butterly (dolomite)	185–206
			Signal Mountain			Signal Mountain	207–208
			Royer			Royer (dolomite)	209–214+
			Fort Sill			Fort Sill	?
			Honey Creek		Timbered Hills group	Honey Creek	
						Cap Mountain	
						Reagan	

strike N. 30° W.), figured 60 feet in thickness to 100 feet measured horizontally as follows:

Section of Honey Creek Formation

Zone	Thickness (feet)	
1	17.96	Irregularly bedded crystalline limestones with much ferruginous material. Much glauconite; weathers irregularly.
2	24.96	Grayish coarsely crystalline limestone with irregular shaly lenses; evenly bedded; much glauconite. Trilobite zones at base and 5 feet above base.
3	7.44	Several crystalline limestone beds exposed. Trilobites at base of zone.
4	19.20	Upper half coarsely crystalline grayish-brown limestones with glauconite; lower half shaly and sandy beds alternating with limestones; trilobites at base.
5	4.38	Crystalline gray limestones; trilobites at base.
6	16.20	Sandstones and limestones alternate in upper part with shaly zone at base resting on coarse Reagan sandstone.
Total	90.14	

The Honey Creek is placed here with the Reagan because of the gradation phase in its lower part. In several outcrops in the Wichita Mountains there is considerable sandstone in the lower part of the Honey Creek and often considerable coarsely crystalline limestone in the upper part of the Reagan. There is a marked difference between the coarsely crystalline Honey Creek and the dense lower beds of the Fort Sill, the basal formation of the Arbuckle group. There is commonly much more glauconite in the Honey Creek than in the beds above, and because of these features they weather differently. The trilobites and brachiopods of the Honey Creek differ greatly from those in the beds above. *Eoorthis remnicha* and *E. wichitaensis* occur in the lower part of the formation about 80 rods to the south of the type section, on the north side of the West Timbered Hills, and in the outcrops on the north edge of the Wichita Mountains. There is often a distinct shaly zone at the top of the Honey Creek separating it from the resistant dense heavy beds of the overlying formation. Accordingly, it is thought best to retain it as a formation in the Timbered Hills group.

In addition to the outcrops adjacent to the two Timbered Hills in the Arbuckle Mountains, excellent exposures of the Honey Creek occur in the Wichita Mountains. Two of these Honey Creek exposures are on the south side of the mountains in the west end of McKenzie Hill, and 2 miles to the west of that hill. Three outcrops occur in the Blue Creek Canyon region north of Mount Scott. The largest exposure is on the northeast side of the canyon, extending in a long narrow strip for 7 miles. Here, also, it attains a thickness of 266 feet. It outcrops along the west edge of the canyon as well, and at the south end

of the canyon, 4 miles north of Mount Scott. These two outcrops are not on the geologic maps of this region. Two other small exposures occur to the northwest, about 12 miles southwest of Carnegie.

ARBUCKLE GROUP

General statement.—The Arbuckle group is here divided into nine formations, four of which are placed in the Upper Cambrian with the three formations of the Timbered Hills group, and five in the Lower Ordovician.

Fort Sill formation.—The name of Fort Sill was given to the lower part of the Arbuckle because of the outcrop in the small quarry worked by prison labor at the south side of the small creek on the east side of the main north–south road through the post, where it consists of limestones, a little shale, and considerable shaly limestone. The exposed edges of beds higher than those in the quarry extend a considerable distance up the hill to the south.

This formation was named and defined by Ulrich (1932, p. 744-745), and two sections were presented where the thickness given is 150 feet or three times as much as that shown in the recent table (Ulrich and Cooper, 1938, Pl. 58, no. 2). The section near the west end of McKenzie Hill, to which he gives a thickness of 150 feet, is designated as the type for this formation. The base of the Fort Sill rests on the Honey Creek formation which extends beneath it 202 feet horizontally to the northwest edge of the exposure; and the top of the Fort Sill extends east of the road south of Signal Mountain, to the base of a fine pebbly brown conglomerate. Frederickson and the writer found that on some well-exposed bedding planes in several small quarries, some of which have been opened recently, the dip was 13 degrees instead of 10 degrees as used in the earlier computations. This formation aggregates a thickness of 223.8 feet or about 74 feet more than was assigned to it before. It consists chiefly of limestones but includes some limestone conglomerates. The formation has sometimes been called the "sponge beds" because, at least locally, siliceous sponge spicules are very abundant. Also, this has more recently been divided into seven zones as compared to four of Ulrich. *Billingsella corrugata* Ulrich and Cooper (1938, p. 73, 74, Pl. 7A) occurs in this type section as well as in the quarry at Fort Sill. Several trilobite zones, some gastropods, and one oölitic horizon are to be noted. The more detailed section will appear soon in a publication by the Oklahoma Geological Survey.

Royer formation.—This formation was named for the old Royer ranch on the west side of the East Timbered Hills, south of Honey

Creek, in the Arbuckle Mountains, where the dolomites are well exposed, and is designated as the type section of the Royer. It was named and defined and assigned a thickness of 100 to 600 feet by Ulrich (1932, p. 744). Recent studies show that it includes six zones of dolomites aggregating a thickness of 680 feet. Five of these zones are separated by four zones of limestone, totalling a little over 90 feet, giving a total thickness of 771 feet for the formation. Most of the dolomite is coarsely crystalline, and some of it is changed to marble. While the color varies greatly, gray and pink predominate. In the Wichita Mountains this formation has been recognized only on the north side in two outcrops 10 to 12 miles south and southwest of Carnegie where it is about 200 feet thick.

Signal Mountain formation.—The type section of this formation is exposed on the west slope of McKenzie Hill, a mile south of Signal Mountain, which is a high igneous hill in the western part of the Fort Sill Military Reservation. The Signal Mountain formation extends from a point a short distance east of the road between secs. 7 and 8, T. 2 N., R. 12 W., eastward up McKenzie Hill to the base of the coarse conglomerates in the basal part of the McKenzie Hill formation. The fine conglomerate at the base of the formation contains small round brownish pebbles, and above this numerouse coarse limestone conglomerates alternate with wide zones of thin-bedded limestones. It includes many fossiliferous zones. In the Arbuckle Mountains it conists of limestones 404 to 478 feet in thickness which lie between two major dolomites—the Royer below and the Butterly above. On the east side of U. S. Highway 77 on the Chapman ranch, the upper part of the formation contains many trilobite horizons and a few brachiopods and gastropods.

This formation was named by Ulrich (1932, p. 746-747) and was divided into five faunal zones aggregating 235 feet in thickness. Frederickson and the writer measured this section again, noting the higher dip, and collected from it in greater detail than ever before. The new measurement of the formation gives a thickness of 290.46 feet divided into 10 zones. Many trilobite horizons occur in the limestones and in the numerous limestone conglomerates which lie between the zones of thin limestones.

Butterly formation.—This formation is named from the ranches southeast of the east end of the West Timbered Hills in secs. 28 and 33, T. 1 S., R. 1 E., owned by John Butterly and the Butterly brothers. However, the best exposures of the formation are to the southeast on the west side of the East Timbered Hills and southeast of the Chapman ranch

buildings on the east side of U. S. Highway 77. The type section is taken on the Chapman ranch where it constitutes the uppermost dolomite of that region. It is exposed also about 3 miles to the east on the northeast limb of the Arbuckle anticline. At the type locality the formation consists of 286 feet as measured by the writer at an earlier date and includes zones 185 to 206 of the published section (Decker and Merritt, 1928, p. 34, 35). A paced measurement on the west side of the East Timbered Hills gave a thickness of 264 feet. Most of the dolomite is coarsely crystalline, and its upper part contains many large quartz grains and some thin zones of fine quartz conglomerates. This dolomite is present in the section east of the Nebo store southwest of the village of Mill Creek, but its limits cannot be definitely placed because of the extensive soil covering.

No outcrops of this formation have been identified in the Wichita Mountains, and it is thought not to exist there, unless it is represented by some of the numerous dolomites at the northwest end of the mountains or by limestones stratigraphically equivalent in some of the other sections.

McKenzie Hill formation.—The type section is exposed near the top of the western slope of McKenzie Hill a mile southeast of Signal Mountain, about 7 miles northwest of Lawton, where it includes 223 feet of limestones and limestone conglomerate divided into eight zones. A more general subdivision into six zones is given below. Limestone conglomerates at the base, 8 feet in thickness, are followed by 27 feet of thin limestones and thicker limestones alternating with limestone conglomerates and containing six horizons of trilobites. The next higher zone contains mostly dense limestones and a few conglomerates aggregating 76 feet in thickness. These are followed by 37 feet of quarry beds consisting of even-bedded limestones, some of which are carbonaceous and some shaly with many fucoids present. These carbonaceous shaly beds contain the lower Arbuckle graptolite zone. Next above is 31 feet of thick beds of limestone with cylindrical cherts followed by 44 feet of thin badly weathered limestones with many fucoids. Thus the formation extends from the base of the conglomerates on the hill slope to the base of the dolomite which caps the hill. This formation is exposed in Quarry Hill, east of the Fort Sill railway station, in the Dolese quarry at Richards Spur, and in all of these quarries the lower Arbuckle graptolite zone has been traced.

In the Arbuckle Mountains this formation extends from the top of the Butterly formation, the uppermost dolomite, for 40 feet above the lower Arbuckle graptolite zone, giving a total thickness of 1122 feet.

The McKenzie Hill formation is here divided into two members—the Chapman ranch below and the McMichel above. The Chapman ranch member includes thick beds of limestone lying directly above the Butterly dolomite. Bridge (1936, p. 983) assigned a thickness of 144 feet to those beds from which gastropods and cephalopods were collected which he considers identical with forms in the Van Buren of Missouri. The McMichel member is named for the ranch southwest of McKenzie Hill, in sec. 14, T. 2 N., R. 13 W., on which this member is well exposed in a quarry and in a ridge extending north of it. In both systems of mountains, heavy beds of limestone alternate with zones of thin-bedded limestones which in parts are shaly and carbonaceous with many fucoids present, among which are the dendroid graptolites of the lower Arbuckle horizon. Also, other fossiliferous zones occur both above and below the graptolites.

Strange formation.—This name is proposed for beds exposed in only four outcrops on the south side of the Wichita Mountains. The name is taken from the Strange ranch on which two of the outcrops occur, in secs. 8 and 9, T. 2 N., R. 12 W. However, the type section of dolomite caps the top of McKenzie Hill in sec. 8 and extends down its southeast slope. A small one lies only a short distance east of McKenzie Hill, and two others lie a little over a mile to the southeast. It is about 80 feet thick at the top of the hill, and about 60 feet is exposed in one of the hills to the southeast. The maximum here is less than one-third the 270 feet assigned to the formation by Ulrich (Ulrich and Cooper, 1938, Pl. 58, no. 4). While the nature of the outcrop is such as to make accurate measurement difficult, this latter figure seems entirely too great. It is a very resistant, gray to pink, mostly coarsely crystalline dolomite. This is the youngest dolomite of importance in either the Arbuckle or Wichita mountains, unless it is the equivalent of some of those south of Gotebo, where numerous dolomites occur, whose equivalents have not been determined. The dolomites of the Strange formation are readily available for quarrying. They are only 6 to 7 miles from Lawton, near good country roads, and are only 1 to 2½ miles from the paved highway.

Cool Creek formation.—This name is proposed for beds that are well exposed in the large Arbuckle limestone outcrop on the Kindblade ranch, about 12 miles southwest of Carnegie. This is the type section and it lies about the middle of the larger ridge on whose crest and sides it crops out. However, the name is taken from Cool Creek on the south side of the Arbuckle Mountains, east of U. S. Highway 77, where the headwaters of the creek cross the rocks of this formation. It has a thickness of 1015 feet at the type locality, between the main *Ceratopea* zone above and the

heavier beds of limestone at the top of the McMichel member of the McKenzie Hill below. It consists chiefly of relatively thin beds of limestone with much chert, part of which is oölitic, and toward the base of the formation the chert is black. In the type locality the rocks of this formation erode more rapidly than those above or below. Similarly, it is relatively nonresistant in the Arbuckle Mountains where it is 1687 feet thick and contains much oölitic chert. Much of it is covered with soil and grass. While it is well developed in the large Arbuckle limestone outcrops on the north side of the Wichita Mountains, it does not outcrop on the south or east side of those mountains, as only rocks of lower horizons are exposed there. It seems to be relatively unfossiliferous except for several marked algal zones.

Kindblade formation.—This formation takes its name from the ranch along whose east side it is exposed, on the north side of the Wichita Mountains, 10½ miles southwest of Carnegie. At that locality it includes about 957 feet of the uppermost outcropping Arbuckle beds, through which the main zone of ceratopeas extends.

In the Arbuckle Mountains it has a thickness of 1216 feet, the top of it being a little over 1600 feet below the top of the Arbuckle limestone. This greater thickness may be due either to the thickening of the beds eastward or to an expansion of the main *Ceratopea* zone. In both mountain regions the rocks of this formation consist chiefly of dark-gray limestones arranged in groups of beds in which one or two thick ones alternate with zones of thinner beds. In general the beds become thicker and more resistant toward the base of the formation, and the chert nodules are also more abundant in the lower part. In addition to the ceratopeas, several brachiopods and gastropods occur which serve to correlate a part of this formation with the Newala limestone of the Southern Appalachian region. Three of the forms common in the Kindblade formation have been illustrated by Butts (1926, Pls. 17, 18) from the Newala limestone of Alabama. They include *Ceratopea keithi, Orospira bigranosa,* and *Hormotoma artemesia.*

Three of the brachiopods from the Kindblade have been described and illustrated by Ulrich and Cooper (1938, p. 162, 169, Pl. 31A, 33B, and 34A). These include *Tritoechia delicatula, T. typica,* and *T. subaequiradiata.*

West Spring Creek formation.—This formation is named for the creek at the west edge of the Arbuckle Mountains, east of Pooleville, where it is well exposed. However, the type section is on the west side of U. S. Highway 77, a little over 3 miles north of Springer. It extends from the top of the main *Ceratopea* zone to the basal conglomerate of the Joins formation at the base of the Simpson group.

In this formation many thin beds alternate with thick beds, and in the lower part of the formation the alternation of these highly tilted resistant

beds with zones of thin shaly limestones has produced the noteworthy "graveyard" topography so highly developed along the hillside on U. S. Highway 77. A thickness of a little over 1600 feet has been assigned to this formation in the Arbuckle Mountains, and it is absent from most of the outcrops in the Wichita Mountains. In fact, rocks of the Arbuckle limestone as young as these have been discovered in only one locality in those mountains. They are exposed in a highway cut at the end of a low ridge 14 miles south and 2 miles east of Mountain View. At that locality only about 250 feet of beds belonging to the middle of the formation is exposed. Their age is determined by a tiny graptolite, *Didymograptus protobifidus praecursor*, which occurs 845 feet below the top of the formation on Highway 77. This variety was named and described from Norway by Astrid Monsen (1936). Other fossils of the formation include the noteworthy gastropod and algal beds near the top and ostracodes and brachiopods at a number of horizons.

Six brachiopods from this formation have been recently described and illustrated by Ulrich and Cooper (1938, p. 154, 165, 172, 173, 174, Pls. 28E, 30C, 34B, 34E, 35B, and 35C). They include *Diparelasma transversum, Tritoechia planodorsata, Pomatotrema magnum, P. murale, P. oklahomense* and *P. transversum.*

The main *Didymograptus protobifidus* zone occurs 808 feet below the top of this formation on U. S. Highway 77 and on the east side of the Washita River below Big Canyon as well as in both of the Crusher quarries north of Big Canyon station. This species was differentiated and described by Elles (1933, p. 98). It was first recognized in North America by the writer (Decker, 1936, p. 304, 1255) from this West Spring Creek formation in Oklahoma. He recognized this species from the Marathon region of southwest Texas and from the limestone near Smithville in northeastern Arkansas and believes that, from the association and description by Ruedemann (1905, p. 690), it occurs in the Deepkill of eastern New York. Beside this occurrence in various parts of America and Europe, Harris (1935, p. 333) has listed this species from the Bendigo region of Victoria, Australia. Thus the middle part of the West Spring Creek formation is connected by means of the graptolite *Didymograptus protobifidus* with similar horizons in three different States and with Europe and Australia (Arbuckle group, total maximum thickness 7292 feet).

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University of Oklahoma, Norman, Oklahoma.

Manuscript received by the Secretary of the Society, January 31, 1939.
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 1323-1356, 7 FIGS. AUGUST 1, 1939

DRAINAGE EVOLUTION IN THE SOUTHERN APPALACHIANS

BY H. D. THOMPSON

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ABSTRACT

Following the Appalachian revolution and the Palisades disturbance the main drainage divide between the Atlantic Ocean and the interior was not far from the Blue Ridge. How the drainage was reversed and the divide shifted to its present position has long been a major question of Appalachian geomorphology commonly answered by regional superposition.

Difficulties involved in assuming widespread perfect peneplanation or an extended coastal plain cover of the Appalachian Highlands invite an alternative interpretation without recourse to regional superposition. Headwater piracy along the asymmetrical watershed resulting from eastward tilting of the Atlantic border region is believed competent to account for the westward migration of the main drainage divide. Thus the process of drainage reversal has been slow and long-continued rather than abrupt. In the northern part of the Southern Appalachians the relations of streams to rock resistance and structure, the characteristics of dry gaps, the drainage of canoe-shaped valleys, examples of recent and imminent piracy, and the present asymmetrical position of the drainage divide support this interpretation.

INTRODUCTION

Fifty years ago Davis (1889) proposed that the southeast-flowing rivers of Pennsylvania had developed by the reversal and dismemberment of a master consequent stream. Six years later Willis (1895, p. 190) stated that the main streams of the Atlantic slope had been superposed across the hard rocks of the Appalachians from the floodplain alluvium of the Kittatinny (Schooley) peneplane. The doctrine of superposition from the Schooley peneplane gained considerable currency. In more recent years it is commonly believed that the Schooley peneplane was not of sufficient flatness to allow superposition from its alluvial cover. To obviate this difficulty and to explain the notable adjustment of streams to structure, Johnson (1931) presented the hypothesis of a pre-Schooley superposition of the Atlantic-slope streams from a coastal plain cover. Johnson's hypothesis has gained wide acceptance in recent textbooks (Schuchert and Dunbar, 1933, p. 388-390; Loomis, 1937, p. 146-166; Fenneman, 1938, p. 258-259). A few recent short papers have questioned the theory of superposition—whether Schooley or pre-Schooley (Ruedemann, 1932; Meyerhoff and Olmstead, 1936; Thompson, 1936). The writer believes with others that the problem of Appalachian drainage evolution should be kept open for further study and discussion.

The hypothesis of progressive piracy and local superposition here advanced is an extension of the hypothesis proposed by the writer in a paper on the Hudson River in 1936. Most of the discussion is confined

to the northern part of the Southern Appalachians, where the Potomac, James, and Roanoke rivers cross the Blue Ridge and drain some of the region on the west. No attempt is made at a full systematic description

FIGURE 1.—*Index map of the Southern Appalachians*

or a review of the literature on the area, as given by Wright (1925, 1931, 1934, and 1936) and Fenneman (1938, p. 1-342).

GENERAL SETTING

LOCATION

The Southern Appalachians include that part of the Appalachian Highlands south of the Potomac River. The structure and surface forms trend northeast by north, a little more easterly than the Atlantic Coast, so that northward the Highlands approach the sea. From southeast to

northwest, the principal subdivisions are the Piedmont province, the Blue Ridge province, the Ridge and Valley province, and the Appalachian Plateaus (Fig. 1).

PROVINCES

The Piedmont province, called the Piedmont Plateau, widens from about 35 miles along the Potomac to about 160 miles in southern Virginia and maintains the latter width to its southwestern extremity in east-central Alabama. Pre-Cambrian crystalline rocks underlie the greater part of the Piedmont region; in addition, narrow belts of metamorphosed Cambrian sediments, fossiliferous Ordovician slates, and several elongate basins of Upper Triassic sandstone and shale cut by diabase dikes and sills are found. On the southeast the Piedmont joins the Coastal Plain along the Fall Zone where the older rocks disappear beneath Cretaceous and Tertiary sediments; the northwestern border is the Blue Ridge. Presumably consequent streams follow the general slope southeastward with some local adjustments to weak rocks. The surface of the Piedmont province is commonly described as an uplifted peneplane now being dissected by streams that have intrenched themselves as much as 200 feet. Many hills and ridges, higher and more numerous near the Blue Ridge, rise above the general level.

The Blue Ridge province extends from northwestern Georgia to a locality near Carlisle in southern Pennsylvania. Throughout most of its length northeast of Roanoke, Virginia, it is essentially a main ridge with a flanking ridge on the northwest and numerous spurs on the southeast, varying in width at the base from about 2 miles, at Afton, Virginia, to about 10 miles, near Luray, Virginia. Southwest of Roanoke the province is much wider, attaining a maximum width of about 70 miles in western North Carolina and eastern Tennessee. The rocks are mostly pre-Cambrian crystallines with Lower Cambrian sediments along the northwestern flank. The Blue Ridge has been in general overthrust northwestward toward the Great Valley above which it rises 1000 to 2000 feet. On the southeast it rises 1000 to 3000 feet above the Piedmont in an extremely ragged slope with many sprawling spurs, deep gulfs, and outliers. In many places it is quite impossible to designate exactly where the Blue Ridge ends and the Piedmont plateau begins. Whatever the original cause of demarcation between these two provinces, their common boundary is now clearly the result of long-continued erosion. Although the summit areas of the Blue Ridge are interpreted by some as remnants of the Schooley peneplane, these areas are small and variable in elevation. Southwest of Roanoke, Virginia, the drainage divide between the Atlantic Ocean and the Gulf of Mexico is in the Blue Ridge

province. Northeast of the city of Roanoke, the Roanoke, James, and Potomac rivers cut through the Blue Ridge.

Between the Blue Ridge on the southeast and the Appalachian Plateaus on the northwest the Ridge and Valley province, 50 to 60 miles wide, extends from central Alabama to the St. Lawrence Valley. Folded, and in many places thrust-faulted, Paleozoic rocks, which in general become younger northwestward, underlie the Ridge and Valley province. The southeastern part, where Cambrian and Ordovician limestones and shales outcrop, is in most places a wide valley with minor ridges, known as the Great Valley; a few imposing ridges of resistant sandstone rise above the Valley surface. The northwestern part, where Silurian, Devonian, and Mississippian sandstones are common, has many high linear ridges parallel to the valley trend. The crests of the higher ridges in the Ridge and Valley province are usually interpreted as approximately the level of the Schooley peneplane, and the broader limestone valley floors as the Harrisburg peneplane. South of Chattanooga, Tennessee, the drainage of this province goes directly southward to the Gulf; from Chattanooga to a few miles west of Roanoke, Virginia, the drainage goes by way of the Tennessee and New rivers to the Mississippi and the Gulf; the remaining, northeastern portion drains to the Atlantic.

Throughout much of its length the Appalachian Plateaus province faces the Ridge and Valley province in a bold escarpment called the Cumberland Front in Tennessee and southwestern Virginia, and the Allegheny Front in northern Virginia, Maryland, and Pennsylvania. In New York it is called the Catskill Escarpment. For 50 miles southwest of the New River there is no well-defined scarp, and the topography of the Ridge and Valley province merges into that of the Appalachian Plateaus. This province is a maturely dissected plateau that slopes northwestward and grades into the Interior Lowland. Slightly flexed Carboniferous sediments dip gently northwestward to the structural axis of the Appalachian geosyncline. In contrast to the Blue Ridge and Piedmont provinces, which are widest in the south, the Appalachian Plateaus broaden northward. The extreme northern and northwestern parts of the plateau drain northeastward through the St. Lawrence; the northeastern part is drained southeastward to the Atlantic by the Hudson, Delaware, and Susquehanna rivers; the remaining, larger portion is drained by the Ohio and its tributaries to the Mississippi and the Gulf.

PRESENT DRAINAGE DIVIDE

The Atlantic-gulf divide (Fig. 1) is an irregular, sinuous line that does not conform to any one geomorphic province or structural unit. From southwest to northeast the divide is progressively farther west in the Ap-

palachian Highlands. Southwest of the Roanoke River the divide is in the Blue Ridge province; the Roanoke and New River divide is in the Great Valley; the James and New River divide is farther west in the Ridge and Valley province; the Potomac and Monongahela divide is in the eastern part of the Appalachian Plateau; northeast of the Potomac River the divide is still farther west in the Appalachian Plateau.

From the divide the distance to the Gulf is much greater than to the Atlantic. This disparity increases from southwest to northeast. From the head of the Savannah River the distance to the Gulf is four times as great as to the Atlantic; from the head of the Roanoke River the Gulf is five times as far as the Atlantic; from the head of the James River the Gulf is six times as far away; from the head of the Potomac the Gulf is seven times as far distant as the Atlantic.

This divide has been interpreted as the approximate axis of greatest uplift of the Schooley peneplane (Fenneman, 1938, p. 260). Under such an interpretation the drainage took its southeastward and northwestward courses from this axis and has persisted to the present, with only slight migration of the divide. The general decrease in elevation of the summits in both directions from the divide has been interpreted as the original slope of the uplifted Schooley peneplane. Although Johnson's interpretation calls for a pre-Schooley initiation of southeast drainage across the Appalachian structure, and Wright allows for considerable northwest migration of the divide in post-Harrisburg time, they seem to agree in general with the above interpretation. It is the writer's opinion that the divide has been migrating northwestward for a long time and is in places as much as 100 miles northwest of its original position. The present southeast slope of much of the upland has been converted from a northwest-sloping surface by the headward growth of southeast-flowing streams.

HYPOTHESIS OF REGIONAL SUPERPOSITION

Two versions—Schooley and pre-Schooley—of the hypothesis of regional superposition are now extant. The version of Schooley superposition assumes a very flat alluvium-covered peneplane; the pre-Schooley version assumes a widespread pre-Schooley peneplane covered with coastal plain sediments. Both assumptions are subject to considerable doubt, but neither can now be definitely disproved. It is not the writer's purpose to discuss the relative merits of the two versions of regional superposition; his main thesis is opposed to any method of regional superposition of drainage in the Southern Appalachians.

Johnson's (1931) discussion deals principally with the area north of the Potomac; nevertheless, the principles and general features that he

uses in support of the theory of regional superposition in the north should apply equally in the south, assuming that both regions have had essentially the same history. Likewise, the principles and many of the features here recited should be applicable in the north. As objections to the reversal of northwest consequent drainage, Johnson (1931, p. 33) gives (1) The directness of the course, (2) the great length, and (3) the parallelism of the southeast-flowing streams. These objections may be open to question. From the heads of the Potomac, James, and Roanoke rivers the distances to the sea along the river courses as compared with straight-line distances are nearly twice as great. The statement that rivers have great length has little meaning except by comparison. When the Atlantic-slope rivers are compared with their opponents flowing into the Gulf their great length is not impressive. Any strict definition of parallelism does not apply to the courses of the Potomac, James, and Roanoke rivers, in the sense that they are parallel from source to mouth. Of course the source of any large river cannot be specifically located; many branches gather water from a large area. The principal upper branches of the James (Jackson River) and the Potomac (South Branch) are longitudinal subsequents that flow in opposite directions from the same low divide at Monterey, Virginia. In crossing the Ridge and Valley province the James and the Potomac are separated by as much as 176 miles; after crossing the Blue Ridge they flow toward each other, the James making one long jog and the Potomac two short ones; at Richmond the James is only 55 miles from the Potomac. The James and the Roanoke are 104 miles apart at the heads of their main branches, 20 miles apart at Lynchburg, and 80 miles apart at their mouths. There is a rectangular pattern controlled largely by structure and relative resistance of the formations; the streams either parallel the principal structural units or cross them directly, but this does not suggest superposition. The fact that the departures of these three rivers from direct and parallel courses corresponds so closely with the distribution of relatively weak rocks makes it appear that the structure is the cause of the particular locations of the streams.

Although one of the main objectives of the theory of regional superposition is to explain the supposed lack of adjustment of streams to structure, Johnson (1931, p. 34-36) notes this remarkable adjustment, which he designated modification of consequent drainage, and gives that as a reason for superposition earlier than on the Schooley peneplane. In explaining the high degree of adjustment an additional cycle is of doubtful value for, as Fenneman (1938, p. 197) says, ". . . even though adjustment in one cycle be complete, readjustment is necessary in the next cycle." It would be less complicated to dispense with regional superposition altogether and to explain the adjustment on the basis of head-

ward erosion, with local superposition in the few places where there is not adjustment.

Alignment of water gaps and wind gaps is given by Johnson (1931, p. 36-37), Wright (1934, p. 48-49), and others as evidence of superposition. Alignment of water gaps is not common in the area here discussed. In most cases where a stream crosses a series of strata it does so in a series of jogs. Dry gaps are so numerous that it is not difficult to find some which can be aligned; but, as will be shown for Thornton and New Market gaps, their relative altitudes are in reverse order for a southeast-flowing superposed stream.

The faint slope of ridge crests toward the master gaps reported by Ver Steeg (1930, p. 218) is cited by Johnson (1931, p. 43) as evidence of pre-Schooley superposition. Whether a water gap is made by a superposed stream or by headward growth, the ridge crest should normally slope toward the gap. The stream in the water gap is local base level for the longitudinal subsequents, which in turn are base level for the resequents and obsequents. Since the longitudinal subsequents descend toward the master stream, the resequents and obsequents have greater power to cut into and lower the ridge as it approaches the water gap. The differential lowering within a few million years at most should be sufficient to make a discernible slope toward the water gap. This feature is no justification for assuming either a pre-Schooley or later superposition.

Explanation of the apparent widespread disregard of streams for rock structure has seemed to many to require a theory of superposition. The writer will attempt to show that the present structural relations of the Atlantic-slope streams in the folded belt of the Southern Appalachians do not require a theory of regional superposition, and he will give other evidence in support of the alternative hypothesis of progressive piracy and local superposition.

HYPOTHESIS OF PROGRESSIVE PIRACY AND LOCAL SUPERPOSITION

ORIGINAL DRAINAGE DIVIDE

In attempting to locate the original main northwest-southeast drainage divide the question arises as to just what particular event, or episode, or time, should be regarded as the starting point. There seems to be no doubt that the divide between the Atlantic Ocean and the Appalachian geosyncline for most of Paleozoic time was somewhere in Appalachia, a part of which is probably represented now by the Piedmont province and the Blue Ridge. Following the Appalachian Revolution the divide was apparently in the region of the Blue Ridge. Overthrust faults west of the Blue Ridge indicate that this old land was elevated more than the

geosynclinal area to the west. Keith (1828) reports that ". . . . these overthrusts descend toward or into the great Blue Ridge anticline which exposes the most ancient rocks and forms the chief line of uplift of the Appalachians." During Triassic sedimentation the highest mountains were west of the basins of deposition (Roberts, 1928, p. 156). After Triassic deposition the formation of the Catoctin Border fault and the normal block faulting of the sediments (Roberts, 1928, p. 163-164) further depressed the region east of the Blue Ridge and probably more clearly defined the Blue Ridge anticlinorium as the main drainage divide. In the absence of more specific data we might place the late Triassic or early Jurassic divide midway between the Catoctin Border fault and the Blue Ridge thrust fault (Fig. 1). Since late Triassic we have no preserved records of folding, faulting, or sedimentary accumulation in the Appalachian Highlands.

EXTENT OF MIGRATION

Assuming that the position of the original divide above noted is approximately correct, the distance between the original and the present divide is progressively greater northeastward. In northwestern Georgia the headwaters of the Altamaha River, opposed by the Chattahoochee flowing directly southward into the Gulf, have probably pushed the divide northwest not more than 10 to 20 miles; the Savannah and Santee rivers, opposed by the Tennessee and New rivers, have cut back 20 to 40 miles; the Roanoke and James rivers, opposed by the New, have pushed northwest 40 to 75 miles; and the Potomac-Monongahela divide is now 80 to 100 miles northwest of its original assumed position. Southwest of Roanoke, Virginia, the divide is still in the crystalline rocks of the Blue Ridge province; northeast of Roanoke, the Roanoke, James, and Potomac rivers have cut through the Blue Ridge and invaded the Ridge and Valley province. The Potomac has crossed the Ridge and Valley province and is now cutting into the Appalachian Plateau. The problem is to determine why, how, and when the divide shifted to its present position.

CAUSES OF MIGRATION

The fundamental cause of the northwestward migration of the divide was and is the shorter distance to the Atlantic than to the Gulf. According to Schuchert's (1933) maps the Atlantic-slope streams originally had some advantage of shorter distance. As shown by these maps, and supported by recent geophysical exploration (Miller, 1937, p. 807-811), the history of Appalachia during most of Mesozoic and Cenozoic time is dominantly that of eastward tilting and subsidence of the eastern part. Repeated eastward tilting has shortened and steepened the At-

lantic-slope streams and renewed their advantage over those flowing in the opposite direction.

As a result of the more pronounced eastward tilting of the Atlantic border north of Cape Hatteras the Potomac, James, and Roanoke rivers

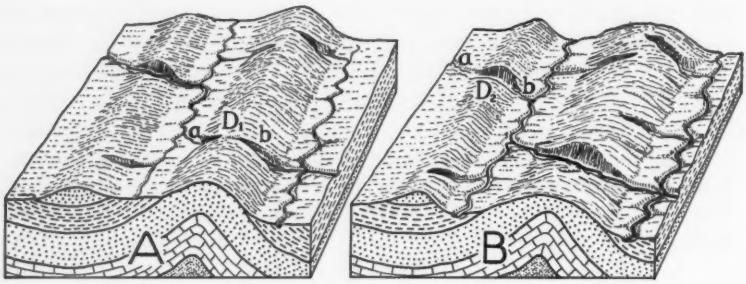

FIGURE 2.—*Diagram showing stream capture and migration of a divide*

The graded gradient of stream b is several hundred feet lower than that of stream a. By progressively undercutting the headwaters of stream a, stream b has pushed the divide from D_1 in A to D_2 in B.

have pushed the divide much farther northwestward than have the streams south of the Roanoke. An additional reason for the greater progress of the northern streams in shifting the divide is that they probably encountered less resistant rock than did the southern streams.

METHOD OF MIGRATION

The writer believes that the divide has slowly migrated northwestward in a normal way by headward erosion and piracy (Fig. 2). Resistant rock barriers have impeded but not prevented the process. The process of migrating drainage by headwater erosion when a divide is asymmetrical has long been recognized but has not been sufficiently emphasized in explaining Appalachian drainage evolution. Probably no one will question the efficacy of this process to accomplish drainage diversion on a small scale. Undoubted examples are not difficult to find (Tarr, 1918, p. 563). Objection may be raised against such large-scale results of the process as are here required. The idea is not new nor does it seem unreasonable to the writer. In discussing river piracy and migration of divides in mountains, Tarr (1918, p. 566) suggested such large-scale migration when he wrote:

"There is thus ample reason for the migration of divides. . . . With migration of divides comes opportunity for river piracy, and it is conceivable that, under favouring conditions, divides may slowly march onward, crossing ridge after ridge even to the opposite side of the mountains, as seems to have been the case in the Appalachian Mountains."

There has been ample time, even though favorable conditions were probably more or less intermittent. If we consider the length of time since the beginning of the Jurassic period as 155 million years (Born, 1933, p. 171) and the maximum northwestward migration of the Potomac watershed as 100 miles, the amount of greatest shifting in this region averages less than seven-tenths of a mile per million years. Some might argue that when a divide becomes located on a resistant rock barrier the process of shifting by headward erosion ceases. No rock is absolutely immune to erosion. Resistant rock barriers slow up the process but do not stop it.

It is not to be thought that the Atlantic-slope streams successively cut through high ridges and diverted large streams from lowlands beyond. The region west of the Blue Ridge originally drained westward and could not be lowered faster than the long-flowing streams could deepen their valleys. From the drainage divide a steeper slope faced southeastward and a gentler slope northwestward, as is now the case in the Blue Ridge southwest of Roanoke. Even where subsequent streams developed along the structural belts, they were high-lying and susceptible to undercutting from the southeast, as is now seen where the Roanoke is undercutting some of the tributaries of the New. Foot by foot the little tributaries of the southeast-flowing streams undercut the little tributaries of their opponents. Each capture strengthened and lengthened the pirate and weakened and shortened the victim. No doubt many captures were simply reversals in valleys partly made by the victims; other captures were made by flank attack. Recent and imminent examples of both types of piracy can be seen along the present watersheds.

In this method of drainage evolution it is expectable that the invading streams would follow the courses of least resistance; however, in downcutting, local superposition on hard rock would be inevitable. As Fenneman (1938, p. 259-260) has aptly said, "There is no such thing as final adjustment of drainage to structure unless hard and soft rocks are separated by vertical surfaces." A stream so superposed might persist across the hard rock for some time, depending on its own volume and gradient in relation to those of its neighbors and on the degree of resistance of the hard rock. If one stream or tributary encountered exceptional resistance its neighbors would have forged ahead and led the drainage around the barrier, providing it was not too far around the barrier.

No attempt is here made to determine the number and positions of peneplanes. The hypothesis expounded in this paper is not dependent on the existence of any peneplane nor is it made invalid by any number of peneplanes. It might well be that several peneplanes, or near-peneplanes, were formed and rejuvenated during the long time and complex

process of drainage shifting. On the other hand, the region may never have been completely base-leveled. Regardless of the number of cycles and the amounts of uplifts it is reasonable to infer that the warping movements have on the average given the advantage of steeper gradients to the Atlantic-slope streams.

ORIGIN OF THE GAPS

Gaps in the hard rock ridges of the Appalachians are commonly designated as water gaps, wind gaps, and cols. Under the regional superposition hypothesis, water gaps were formed by superposed consequent streams that have survived because of their superior volume or favorable location. Wind gaps were likewise made by the downcutting of superposed consequent streams; they were left dry when the streams that made them were captured by tributaries of the stronger or more favorably located consequents, and the altitudes of their floors show the extent of downcutting before capture. Cols are inconspicuous shallow notches, or sags, made by selective weathering and headwater erosion along weak zones in the ridge-making formations; they were never occupied by through-flowing streams. As both wind gaps and cols are now dry, distinction between them, based on size, shape, and depth, is uncertain.

In the hypothesis of progressive piracy and local superposition, great importance must be assigned to selective weathering and headwater erosion in the formation of gaps. An essential requirement is that a col, by the various processes of weathering and headwater erosion, may be sufficiently deepened to provide opportunity for piracy and thus become a water gap. In this interpretation, probably most of the so-called wind gaps in the Appalachians are deep cols, and many of the water gaps are the result of capture. Some gaps are found where streams are locally superposed on resistant rock, and doubtless there are true wind gaps as a result of piracy. The formation and collapse of solution tunnels may possibly account for some gaps.

EVIDENCE SUPPORTING THE HYPOTHESIS OF PROGRESSIVE PIRACY AND LOCAL SUPERPOSITION

CONFORMITY OF STREAMS TO STRUCTURE

General statement.—According to the hypothesis of regional superposition the large streams crossing the Blue Ridge and the ridges of the Ridge and Valley province were superposed from a surface that concealed the underlying structure until they were firmly intrenched. Such streams are said to be unaffected by present structure and hold their unadjusted courses because of great volume. The hypothesis of progressive piracy holds that these Atlantic-flowing streams have grown headward and in

volume by progressively tapping the northwest drainage and that they owe their present positions to favorable rock structures and short distances to the sea.

By favorable rock structures is here meant that the streams cross the ridges where the ridge-making formations are relatively weak. Adjusted drainage would thus include not only the streams that follow the strike along weak formations but also those that cross the resistant formations along lines of weakness. It is an adjustment controlled by two important factors—the distribution of weak rocks and the distance to base level. Distance to base level prohibits much of the drainage in the Appalachians from following the strike of weak formations to the sea.

Unfortunately for geomorphologists, rock formations are not mapped according to their erosive resistance. There seems to be no satisfactory method of measuring this resistance experimentally. Consequently, one must depend on nature. A formation may be very resistant in one locality and quite weak in another; comparative resistance values in one locality are not necessarily true in another locality. For example, the Catoctin greenstone in Stony Man Mountain, near Luray, is a hard columnar basalt that holds up the surface to more than 4000 feet, and the Cambrian sandstone forms a minor flanking ridge. In the vicinity of the Potomac the Catoctin is a fissile, easily eroded schist in the lowlands, whereas the Cambrian sandstone makes the principal ridges. In this study relative resistance of the rock formations was judged primarily by topographic expression. In each area discussed below, the total area of outcrop of each formation was measured with a planimeter, and likewise the area of outcrop above a certain altitude. From these data the percentage of the total area of each formation lying above a specified altitude was calculated. These relative percentages were taken to represent the relative resistance of the formations. Although other factors such as differential uplift, stratigraphic position, distance from main streams, and possible peneplanes, have some importance, it is believed that topographic expression provides a fair basis for approximating the relative erosive resistance of the rocks in a region that has long been subjected to erosion. Some formations in each locality could thus be selected as having pronounced resistance and others as having pronounced weakness. The relations of some of the main drainage lines in the Southern Appalachians to these hard and weak formations are indicated below.

Potomac River.—At Harpers Ferry, the Potomac crosses the Blue Ridge. The Blue Ridge province here consists of the Blue Ridge of Virginia, which extends across the Potomac for a short distance into Maryland as Elk Ridge, and South Mountain of Maryland, which extends into

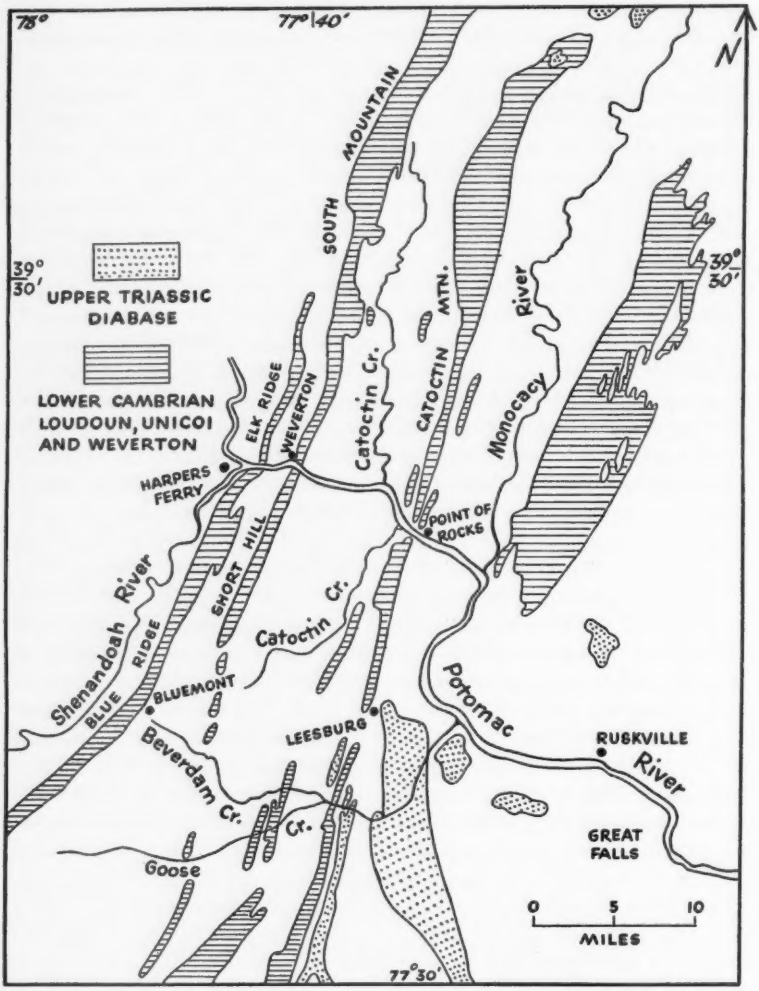

FIGURE 3.—*Geologic map of Potomac gap in the Blue Ridge and vicinity*

northern Virginia as Short Hill. The narrow overlapping ends of these two ridges are separated by a narrow valley, making a total width of about 4 miles for the province. Eight miles to the east, Catoctin Mountain, Hog Back Mountain, and other lesser ridges rise above the general

level in the Piedmont province. Catoctin Mountain is included in the Blue Ridge by some writers.

The Lower Cambrian Loudoun, Unicoi, and Weverton formations are unquestionably the most resistant rocks in this locality. Their hard quartzitic members form the backbones of the Blue Ridge, South, and Catoctin mountains. Not all of the Loudoun is resistant rock, but resistant and weak members are not differentiated on the available maps. Between the ridges the pre-Cambrian rocks have been reduced to lowlands. In the Triassic area, east of Catoctin Mountain, the diabase may be considered as a comparatively resistant formation.

The Potomac, in crossing the Blue Ridge and the western part of the Piedmont province is favored by a narrowing or separation of the resistant rocks (Fig. 3). At Harpers Ferry and at Weverton, the outcrops of the sandstones are very narrow; at Point-of-Rocks, the sandstone appears discontinuous across the river; east of Point-of-Rocks, the Potomac follows the weak Triassic shales. The position of the river here strongly suggests competitive selection instead of chance superposition.

Streams that cross the trend of the formations in the Ridge and Valley province have been described as superposed because they cross the structure instead of following it. It is hardly to be expected that streams should follow indefinitely the structural trend regardless of the general slope of the surface and distance to the sea. Although no quantitative data are available for calculating how far a stream may depart from a direct course in order to avoid a resistant rock formation, the degree of resistance of the barrier no doubt is the principal factor.

From Keyser to Harpers Ferry, West Virginia, the direct distance is 66 miles; by way of the Potomac it is about twice as far. For about half of its course between these two points the Potomac crosses the strike of the rocks; in the other half it parallels the structure. In this part of the Ridge and Valley province the Tuscarora and Pocono sandstones are the most resistant formations. Immediately west of Keyser the Pottsville, another resistant formation, crops out in the Appalachian Plateau. The Oriskany sandstone, although a ridge-maker in parts of this area, is of variable composition and resistance and is not included with the persistently hard formations. The map (Fig. 4) shows that the Potomac avoids the Tuscarora and the Pocono throughout the entire width of the Ridge and Valley province. This relation of the river to resistant rocks makes it seem improbable that its course is due to the fortuitous circumstance of superposition.

Wright (1934, p. 46-47) mentions the Potomac gap in Knobly Mountain at Cumberland, Maryland, as especially indicative of superposition because this mountain is anticlinal. The writer cannot see how either an

anticlinal or a synclinal structure could block a headward-growing stream. When the main divide was located at Knobly Mountain, as it probably was at some time in its westward march, the topographic expression at this locality was somewhat different from that of the present because of

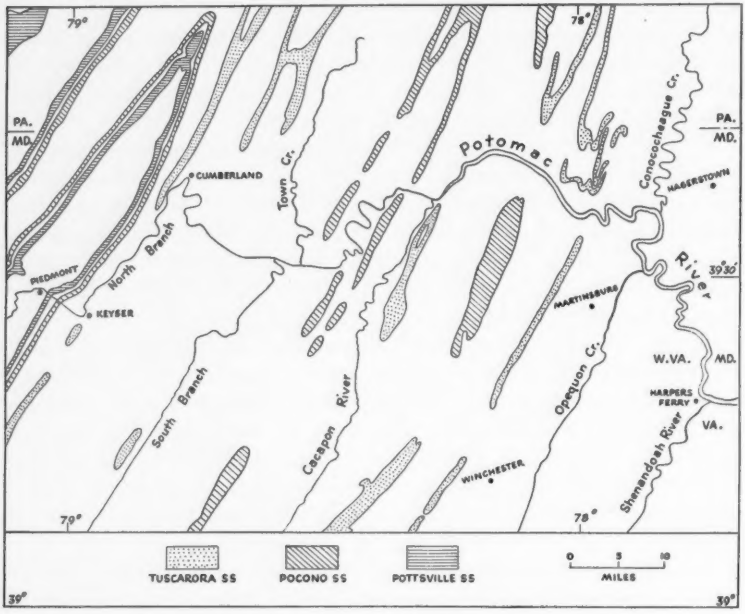

FIGURE 4.—*Geologic map of Potomac basin west of the Blue Ridge*

subsequent erosion. The essential difference was that the grade of the streams west of the divide was probably several hundred feet higher than that of the streams on the east. Probably strike streams developed on the weak rock on each side of the divide, and their tributaries gnawed into the dividing ridge, forming cols where closely spaced joints or other weakness existed. Eventually a col was lowered until the high-lying drainage was diverted to the east through a water gap. After the divide shifted westward the weak rock area west of the former divide was lowered nearly to the level of that on the east. Probably in this manner the drainage divide has repeatedly shifted westward (Fig. 2).

Knobly Mountain at this locality consists principally of the Oriskany sandstone. The northeast pitch carries the Tuscarora below the surface.

Further downcutting will expose this hard formation, across which the river will be, at least temporarily, locally superposed. The fact that the Potomac crosses the Ridge and Valley province without traversing the Tuscarora or the Pocono is considered as evidence favoring the hypothesis of progressive piracy rather than regional superposition.

James River.—Between the Potomac and the James, a distance of 158 miles, there is no water gap in the Blue Ridge. At Balcony Falls the James leaves the Appalachian Valley, cuts through the Blue Ridge and across the Piedmont province to Lynchburg. At Lynchburg the river turns abruptly northeastward and follows the outcrops of Cockeysville marble and Newark sediments for 43 miles before again turning southeast. Of the three water gaps in the Blue Ridge of northern Virginia, the James has the nearest appearance of having been superposed, since it crosses the greatest total thickness of apparently resistant rocks. However, close inspection of the geologic map (Nelson, 1928) strongly suggests that even this river has been guided through the Blue Ridge by the path of least resistance within many miles. Southwestward for a distance of 20 miles no other route appears so easy; northeastward for at least 40 miles no other route appears to offer as little resistance as the course followed by the James through the Blue Ridge.

Using the criterion of percentage of total area above 2500 feet in elevation to evaluate the relative resistance of the rocks, it is found that in that part of the Blue Ridge from a point 20 miles southwest of the Roanoke to a point 20 miles northeast of the James, a distance of 76 miles, the hypersthene granodiorite is the most resistant, the Cambrian sandstones (Unicoi and Erwin) second, and the Marshall granite third. No other rocks outcrop above 2500 feet. Figure 5 shows that the outcrops of these resistant rocks are relatively narrow where the James cuts through the Blue Ridge at Glasgow. This fact suggests that the course of the James may have been determined by competitive erosion rather than by chance superposition.

Another factor that may have been of importance in fixing the location of the James gap is the presence of dikes. Just below the Bedford power plant dam at the southeastern entrance to the gorge most of the river bed can be examined at low water. Eight dikes of a basic composition, totalling 67 feet in thickness and paralleling the river, were observed here. There may be more, as the deeper parts of the channel could not be examined. Because of dams and deep water the probable continuation of these dikes upstream through the gorge could not be observed.

From Covington, Virginia, to Glasgow, the James crosses the trend of the structure in the Ridge and Valley province. The direct distance be-

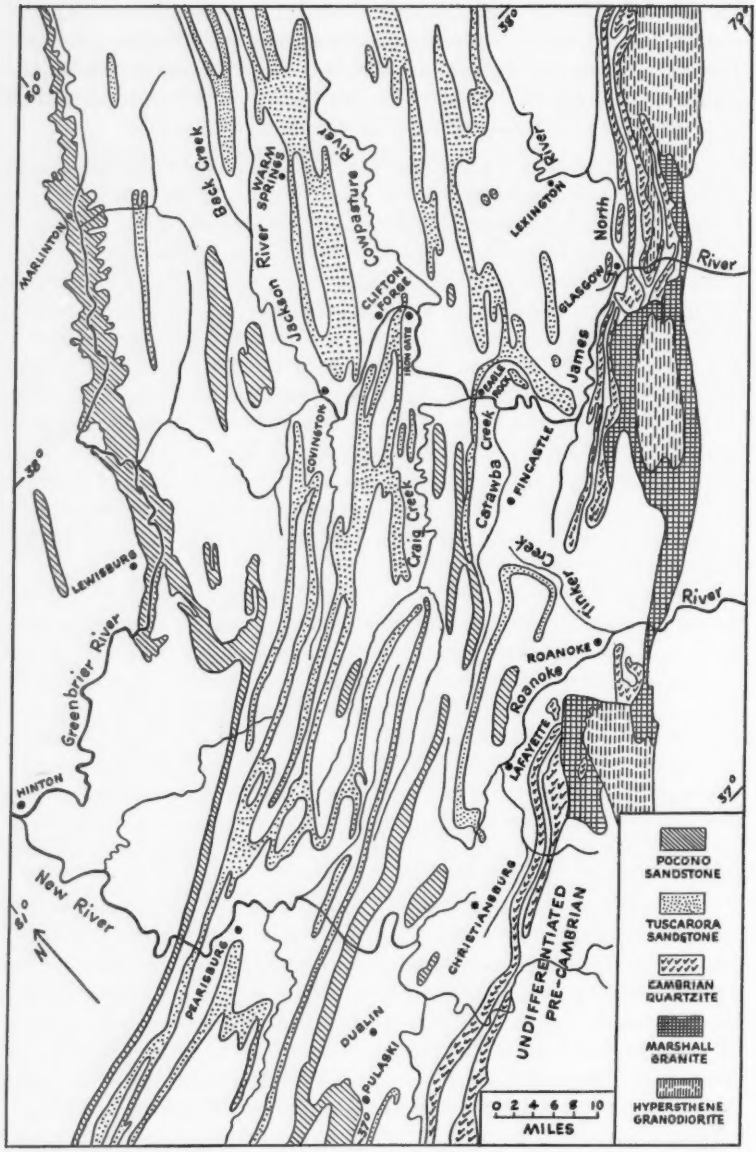

FIGURE 5.—*Geologic map of part of the James, Roanoke, and New River basins*

tween these two points is 32½ miles; by way of the river 76 miles. It is expectable that a river will wander regardless of the method of origin. If this river were superposed its manner of turning aside would not necessarily have any relation to the present distribution of rocks. On the other hand, if the river has invaded this area by headward erosion, its departures from a direct course would be to follow the weaker rocks.

The Tuscarora and Pocono sandstones are the resistant formations in this area. Their distribution (Fig. 5) is such that it would be impossible for a stream to avoid them entirely. If a straight line were followed between Covington and Glasgow, the river would cross 6½ miles of Tuscarora sandstone; by turning as it does it crosses less than a mile of that hard rock.

Where the James crosses the Tuscarora, at Eagle Rock and at Iron Gate, it is said to be superposed (Wright, 1934, p. 49-50). At Eagle Rock, the map shows, it could not have done better; both to the northeast and the southwest the hard rock outcrops are broader. At Iron Gate the river could have avoided the Tuscarora by swinging farther to the northeast. Two possible explanations may be offered for this seemingly anomalous course. First, it may be that the law of diminishing vigor with increasing distance is the correct explanation. There is such a law, no doubt, but it has never been quantitatively expressed. In that case, the stream found it easier to take a short cut across this hard rock where it narrows suddenly than to work entirely around it. The other possible explanation is that of local superposition. The anticline of Rich Patch Mountain pitches northeastward; and the Tuscarora sandstone is overlain by weaker rocks which now form the lowland followed by Cowpasture River. Probably the original valley across this anticline was made before the Tuscarora was uncovered, and the stream has persisted in spite of the resistant rock encountered in downcutting.

Such local superposition is entirely different from the regional superposition postulated by previous writers. One might expect to find many cases in the Ridge and Valley province where streams are temporarily out of adjustment on recently exhumed hard rocks. But it is not likely that such maladjustments will long continue in competition with streams favored by both shorter distance and weaker rocks.

In the Ridge and Valley province, the northeast-flowing tributaries of the Potomac are longer than the southwest-flowing tributaries of the James. Wright (1934, p. 50) says that differential uplift at the close of the Schooley, and perhaps also at the close of the Harrisburg, cycle is the probable explanation of this asymmetrical drainage divide. From the divide, the distance to the sea by way of the James is approximately the same as by way of the Potomac. Thus, with respect to distance to the

sea, the divide is not asymmetrical and apparently requires no theory of warping or other special explanation. Since no resistant barriers separate these opposing tributaries they have become well adjusted with respect to the distance to base level.

Roanoke River.—The Roanoke Gap in the Blue Ridge only slightly resembles a gorge. Most of the resistant rocks of the Blue Ridge are absent for a distance of 10 miles. Probably no one would question that this gap is located at a place of pronounced weakness in the Blue Ridge. Although Wright (1934, p. 46) classifies the James as superposed across the Blue Ridge, he finds no evidence of superposition of the Roanoke.

The Roanoke River drains only a comparatively small portion of the southeastern part of the Ridge and Valley province. This is probably caused by the much longer distance to the sea from the Blue Ridge by way of the Roanoke than by way of the James or the Potomac. Also, the westward drainage by way of the New River is more direct than it is from the James-Greenbrier divide.

The map (Fig. 5) shows that in working headward into the Ridge and Valley province the Roanoke has avoided the resistant rocks. As in the James basin, the hard formations are the Tuscarora (Clinch) and Pocono sandstones. In only one place does a tributary of the Roanoke cross one of these formations. Between Christiansburg and Lafayette the North Fork cuts across a narrow band of the Tuscarora sandstone near the end of its outcrop. This is probably a case of local superposition from the overlying weak Devonian rocks.

Other southeast-flowing streams.—None of the Atlantic-slope streams southwest of the Roanoke cut through the Blue Ridge province. The Peedee, Santee, Savannah, and Altamaha rivers as compared with the Roanoke, James, and Potomac have greater distance to the sea, a wider area of crystalline rocks in the Piedmont and Blue Ridge provinces to retard their headward growth, and shorter Gulf-slope streams opposing them.

Summary.—The rivers that cross the Blue Ridge and Ridge and Valley provinces do so where the hard rock outcrops are relatively narrow and the weak rock outcrops are relatively broad. Two possible conclusions may be drawn from this fact: (1) Many streams were superposed across this belt, and only those most favorably located have persisted in their courses; and (2) many streams have worked northwestward from the original divide, and those most favorably located have been most successful. Consideration of additional evidence should help us in choosing between the two possibilities.

DRY GAPS IN THE BLUE RIDGE OF NORTHERN VIRGINIA

General statement.—The term "wind gap" is now commonly restricted to abandoned water gaps, while those gaps not believed to have been former water gaps are designated as "cols." The writer finds no reliable

TABLE 1.—*Tabulation of gaps in the Blue Ridge of northern Virginia according to depth*

Depth below summits (In feet)	Number of gaps	Depth below summits (In feet)	Number of gaps
100–199	41	1200–1299	2
200–299	38	1300–1399	1
300–399	29	1400–1499	1
400–499	17	1500–1599	0
500–599	6	1600–1699	0
600–699	9	1700–1799	0
700–799	3	1800–1899	0
800–899	6 (includes Roanoke Gap)	1900–1999	0
900–999	4	2000–2099	0
1000–1099	4 (includes Potomac Gap)	2100–2199	0
1100–1199	2	2200–2299	1 (James)
		Total	164

criteria on which the gaps in the Blue Ridge of northern Virginia may be separated into wind gaps and cols; he therefore uses the term "dry gap", or "notch", for any gap not now occupied by a through-flowing stream.

The profile of the crest of the Blue Ridge in northern Virginia shows that practically the entire crest consists of gaps, summits, and the slopes that lead to them. Flat summits are few in number and small in area. Including all notches 100 feet or more in depth and measuring them as accurately as possible on the available maps, 161 dry gaps and 3 water gaps can be counted in the distance of a little more than 200 miles from Harpers Ferry to Roanoke (Table 1).

Those who advocate the theory of regional superposition designate the more prominent of these notches as wind gaps, whose southeast-flowing streams were diverted by the development in the Great Valley of longitudinal subsequent tributaries of the master streams. Snickers Gap, about 500 feet deep, is cited as a typical wind gap (Willis, 1895, p. 190-192). If depth is the proper criterion, all those dry gaps in the Blue Ridge that are 500 feet or more deep meet the requirement for wind gaps. Characteristics of these gaps discussed below do not support the interpretation that they were formed by southeast-flowing streams.

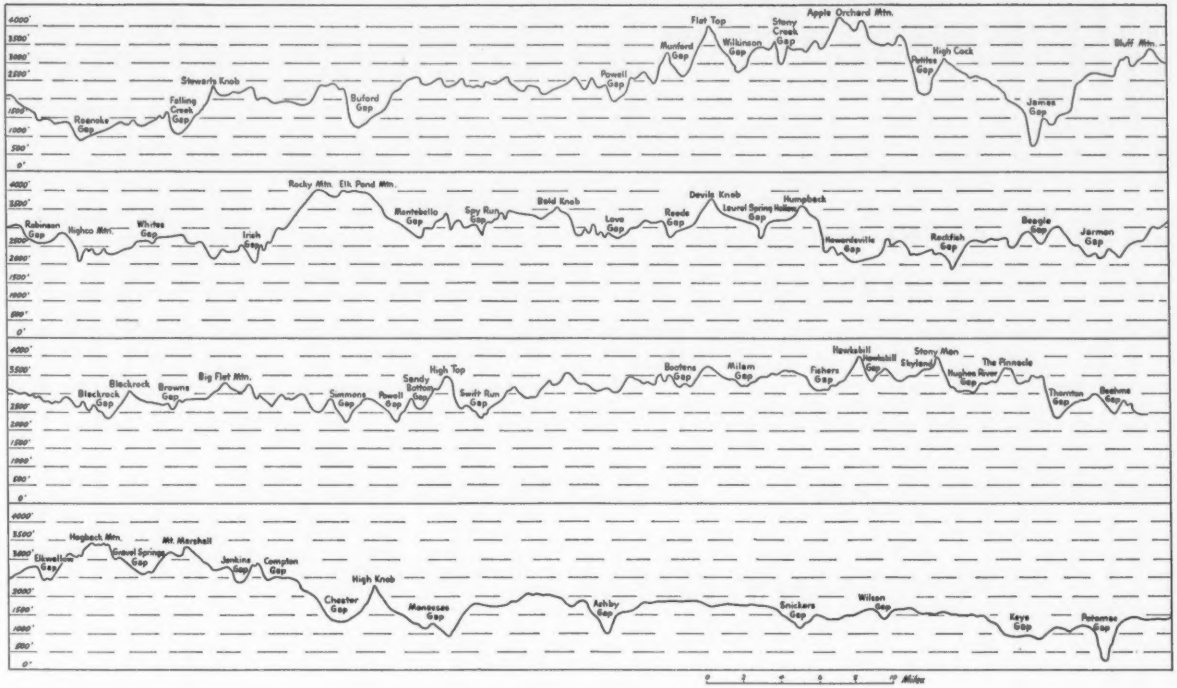

FIGURE 6.—*Profile of the Blue Ridge crestline*

Altitudes and depths of the gaps.—Prominent dry gaps occur at all altitudes from about 850 feet up to 3300 feet. There is no great preponderance at any particular elevation and no regularity of increase or decrease in any direction. On the whole the gaps become higher away from the master streams. The depths of the gap floors below the bounding summits are also highly variable; thirty-six are 500 feet or more deep (Fig. 6).

The altitudinal relationships of the gaps are not as would be expected if they represent progressive captures by tributaries of the three surviving transverse streams. Fenneman (1938, p. 168-171) cites the altitudes of Snickers, Ashby, and Manassas gaps as being consistent with the theory of capture by the headward-growing Shenandoah following uplift of a peneplane. These three gaps are the second, third, and fourth of the prominent gaps southwest from the Potomac. They do have the proper relation. But, the first gap (Keys), 4 miles southwest of the Potomac, is lower than either of the three named, and the fifth (Chester) is higher. On the basis of neither their altitudes nor their depths is it conceivable that all the gaps are related to any one peneplane.

In their paper on drainage changes in the Shenandoah Valley region Watson and Cline (1913) realized that the distribution of large and small gaps in the Blue Ridge was evidence against the capture by the Shenandoah of all the streams that flowed through them, in order from Harpers Ferry southwestward. These authors proposed a more complex succession of captures through two cycles. Beginning in the Tertiary cycle with the capture of all small streams flowing through the shallow gaps by tributaries of the larger streams, three drainage systems developed north of the James, each with an outlet through the Blue Ridge. Rockfish River flowed through Rockfish Gap at Waynesboro, Goose Creek through Manassas Gap, and the Potomac through Harpers Ferry Gap. Later in the same cycle a tributary of Goose Creek captured Rockfish River about Waynesboro, and then most of the Shenandoah Valley drainage went through Manassas Gap. Early in the Shenandoah cycle the young Shenandoah, tributary of the Potomac, captured Goose Creek at Manassas Gap, diverting all the water to the Potomac. The succession of drainage changes proposed by Watson and Cline to account for the great variation in altitudes of the Blue Ridge gaps probably has some advantages over that of having all the captures take place in regular order by tributaries of the present main streams. However, the distribution of deep and shallow gaps adjacent to each of the proposed main drainage systems is argument against the capture of the streams that flowed through them by tributaries of these main streams (Fig. 6). In addition, there would seem to be some difficulty involved in having a large piratical river, such

as Goose Creek must have been when it carried most of the drainage of the Shenandoah Valley, captured by a small tributary of the Potomac. Furthermore, this scheme does not account for the fact that Keys Gap is lower than Manassas Gap. If the gaps have been made by weathering and headwater erosion, it is expectable to find them at various elevations and of various depths; it is also expectable that their floors become higher away from the master streams.

Gradients of streams flowing from the gaps.—In the majority of those gaps more than 500 feet deep the southeast-flowing streams have steeper gradients than those flowing to the northwest. Waterfalls, such as South River Falls (3 miles northeast of Swift Run Gap, on the south half of Shenandoah National Park sheet) and Crabtree Falls (3 miles southeast of Montebello, on the Buena Vista sheet) are more numerous on the southeast than on the northwest. Streams are beheaded because they are less vigorous than their opponents, and by losing their heads their vigor is further reduced. The streams on the southeast slope of the Blue Ridge are now more vigorous than their opponents on the northwest slope, a condition not in harmony with the hypothesis of piracy of southeast-flowing streams. Although the present superior vigor of the streams on the southeast slope of the Blue Ridge is not proof of their past superiority, it is presumptive evidence against the capture of their headwaters on the northwest.

The interpretation here presented is that the streams flowing northwest from the Blue Ridge gaps originally continued northwest. The original Atlantic-Gulf divide was probably not far east of the present crest of the Blue Ridge. First the Potomac, later the James, and finally the Roanoke worked headward through the Blue Ridge and sent out longitudinal subsequents which beheaded the northwest-flowing streams and diverted their headwaters to the Atlantic. The streams on the northwestern slope of the Blue Ridge now have steeper gradients than they did before capture; but their opponents on the southeast still have the advantage of shorter distance and steeper gradient and are slowly pushing the divide northwestward. Some of the present dry gaps in the Blue Ridge may possibly represent cols on the original main divide that have been lowered as the whole mountain crest has been worn down. On the other hand, the shifting of the outcrops during reduction may have been so great that the present notches have little relation to the original cols.

Relation of gaps to structure.—Gaps made by superposed streams need not be adjusted to structure, whereas those made by headward growth should conform to some structural weakness. Inspection of the geologic map (Nelson, 1928) shows some of the broader structural relations of the

gaps in the Blue Ridge. Examination in the field discloses details not shown on the map. The writer has not examined the conditions at all the gaps, but many of those examined show some structural weakness.

Some of the larger gaps are located where the resistant rocks have relatively narrow outcrops. The hypersthene granodiorite and the Cambrian quartzites are the most persistently resistant rocks in the Northern Blue Ridge. From the Potomac southwestward for 40 miles the Catoctin is comparatively weak, whereas it is quite resistant farther southwest. Keys, Snickers, Ashby, Manassas, and Rockfish gaps are all located where the most resistant formations have narrow outcrops. Some gaps are located along the lower contact of the Catoctin greenstone. The lower part of this formation, in some places at least, is vesicular, thinly layered, and weak. The surface on which it was extruded was evidently irregular, or later deformed, so that its contact with the underlying rock is in places steeply inclined. Thornton and Hughes River gaps are of this type. Jarman, Love, Montebello, Irish, Powell (Bedford sheet), and Buford gaps are located at points where the sinuous crestline of the Blue Ridge is intersected by a thrust fault (Stose, 1919, map). These gaps and the coves that lead to them are oblique to the general trend of the mountain. Closer examination in the various gaps might disclose other types of weakness, such as crushed zones, closely spaced joints, and weak dikes. Further investigation of these as well as other gaps in the Appalachians should prove fruitful.

Other characteristics of the gaps.—Instead of the well-defined V-shape, many of the large gaps, such as Manassas, Hughes River, Fishers, Swift Run, Jarman, and Love, have notched floors. In some gaps the separate floors are at about the same altitude; in others they differ as much as 300 feet. Gaps of this type do not suggest abandoned water gaps, unless a considerable amount of subsequent deepening is allowed. If this is allowed, it is no more difficult to allow for the formation of the gaps without the aid of through-flowing streams. Were the gaps made by superposed streams, those in the flanking ridges should correspond in number and position with those in the main, or central, ridge. This is not true in many places. On both sides of the Blue Ridge many coves do not lead to gaps in the main ridge. These coves and notches in the flanking ridges were certainly made by the headward growth of small streams. The water gaps of the Potomac and the James are so steep-sided that, if they were abandoned, talus would accumulate in great quantity on the bottom and lower slopes. As far as has been learned, the dry gaps in the Blue Ridge have very little talus. Profiles do not show talus slopes; and road excavations expose bedrock deeply weathered *in situ*.

ALIGNMENT OF GAPS

About 12 miles west of the Blue Ridge, the synclinal Massanutten Mountain extends 50 miles from Front Royal to a point near Harrisonburg. New Market Gap is the only prominent transverse gap in this mountain. Twelve miles west of New Market, the North Fork of the Shenandoah River cuts through Little North Mountain at Brocks Gap. The alignment of Thornton Gap in the Blue Ridge, New Market Gap in Massanutten Mountain, and Brocks Gap in Little North Mountain has been cited as striking proof of a former superposed southeast-flowing stream (Wright, 1934, p. 48-49). Presumably, the North Fork of the Shenandoah River formerly flowed from Brocks Gap through New Market and Thornton Gaps. Capture of this transverse stream by the headward-growing Shenandoah caused the abandonment of Thornton and New Market gaps. The elevation of Thornton Gap is 2250 to 2300 feet, of New Market Gap 1800 to 1900 feet, and of Brocks Gap 1000 to 1100 feet. Brocks Gap is occupied by the North Fork of the Shenandoah River, which explains its low altitude; but the fact that New Market Gap is about 400 feet lower than Thornton Gap is difficult to explain by the theory of regional southeast superposition. Warping might explain the higher stand of Thornton Gap, but the amount seems too great for such a short distance. According to Watson and Cline (1913, p. 358), the transverse stream was first beheaded by the South Fork of the Shenandoah River in the vicinity of Luray and later by the North Fork at New Market. In the interim between the two piracies the New Market Gap was lowered about 400 feet. This plan could explain the incongruous altitudes of the gaps, but there is no apparent reason why North Fork should have lagged so far behind South Fork in growing headward up the valley.

The difficulties might be removed if we picture a stream flowing northwest from Thornton Gap through New Market Gap. Such a stream, being near the head of the original northwest drainage, might have a gradient of 25 feet per mile, which would account for the difference in elevation between the two gaps. Or, as might be expected of streams pushing the divide westward, the piracy would occur first on the east side of Massanutten Mountain, and further lowering of the New Market Gap would be accomplished before the capture at New Market. Both gaps have no doubt been subsequently lowered by the little streams heading in them.

DRAINAGE OF CANOE-SHAPED VALLEYS

Drainage conditions in some of the simpler anticlinal and synclinal valleys may throw light on the problem of drainage evolution. Burke Garden, in Tazewell County, Virginia, a small anticlinal valley about 10 miles long, is drained through one gap in the western rim; the eastern

rim is unbroken. About 6 miles southwest of Iron Gate, and extending southwesterly for about 10 miles, the Rich Patch anticline is unroofed to make a canoe-shaped valley which drains through two gaps in the western rim. Warm Springs Valley, in the Warm Springs anticline, is drained by six small streams through gaps in the western rim. Bolar Valley (Big Valley), in the Jack Mountain anticline, has two outlets through the western rim. The southern part of Crabbottom Valley, in the Back Creek Mountain anticline, drains westward through Lower and Mill gaps, whereas the northern part of the same valley drains eastward through Vanderpool and Crabbottom gaps. Among the synclinal valleys, Cedar Creek Valley, in Short Hills, west of Natural Bridge, drains eastward through Cedar Creek Gap. Spring Gap, farther south in the same rim, is nearing completion. Little Back Creek Valley, in Back Allegheny Mountain, has an outlet eastward at Mountaingrove. The rather complex synclinal valley in Massanutten Mountain drains eastward through Harshberger, Cub Run, and Passage Creek gaps. Each of the synclinal valleys named is drained through the eastern rim; each of the anticlinal valleys, with the exception of the northern half of Crabbottom Valley, is drained through the western rim. In most cases the water gaps are not matched by dry gaps in the opposite rim.

It is difficult to visualize a method of regional superposition whereby streams in different localities are superposed across only one limb of a fold. It is equally difficult to imagine that superposition would affect only the western limbs of anticlines and the eastern limbs of synclines. Clearly the outlets of these anticlinal and synclinal valleys have been made by headward erosion and continued downcutting. Because of greater thrusting from the east, the western limbs of the anticlines and the eastern limbs of the synclines are usually steeper and have narrower outcrops of resistant rock; consequently the gaps are made in these steeper limbs (Wright, 1925, p. 48). The method of unroofing of anticlines and the formation of gaps in one limb by headward erosion of small streams has long been illustrated in textbooks. If we accept the undisputed fact that the little streams in these cases have made gaps by headward erosion, it should not be difficult to visualize the formation of other gaps, even those of larger streams, in a similar manner.

RECENT AND IMMINENT PIRACY

In the Blue Ridge.—It is well known that the Blue Ridge escarpment southwest of Roanoke is retreating (Davis, 1903). Evidence is here presented that the drainage divide of the northern Blue Ridge is also shifting northwestward. In addition to the fact, already stated, that the majority of little streams on the southeastern slope of the Blue Ridge have steeper

gradients than those on the northwestern slope, a number of the highest summits are southeast of the present drainage divide, and the stream pattern indicates that recent piracies have occurred.

According to the theory of regional superposition the Blue Ridge upland surface formerly sloped southeastward, and many streams followed this southeast slope across the Blue Ridge until they were captured on the northwest by longitudinal subsequent tributaries of the three surviving transverse streams. In cutting narrow valleys into this southeastward-sloping upland from the northwest, it might be expected that some of the interstream uplands now to the northwest of the drainage divide would show the southeast slope and the little northwest-flowing streams should have barbed tributaries, if the piracies occurred not too long ago.

Actually, the reverse is true. Whereas only one locality is found where the highest summit is northwest of the present divide (Knob Mountain, east of Rileyville, in northeastern Page County), one finds at least 10 localities where the highest part of the mountain is southeast of the present drainage divide. Little streams flow northwest some distance before making an acute turn southeastward. Places showing the latter condition are: Fork Mountain, 1 mile south of President's Camp, near the head of Rapidan River; Flat Top, 1 mile southeast of Simmons Gap; Loft Mountain, 1 mile southeast of the Greene-Albemarle-Rockingham County boundary; Cedar Mountain, 1½ miles south of Browns Gap; Bucks Elbow Mountain, 2 miles east of Jarman Gap (all on Shenandoah National Park sheet, south half); Three Ridges, on Buckingham and Buena Vista sheets; Headforemost Mountain, on Bedford sheet.

In the vicinity of Bucks Elbow Mountain and Jarman Gap the features indicated above are probably most striking (Fig. 7). The summit of Bucks Elbow Mountain is 2 miles to the east of and 1000 feet above Jarman Gap. A small stream flows down the western slope of Bucks Elbow Mountain toward Jarman Gap in a shallow valley directly in line with Sawmill Run, which flows westward from the gap to the Shenandoah River. Just before reaching the gap, the small stream quickens its fall and enters the steep-sided valley of the South Fork of Moormans River, which flows northeast 3 miles to join the North Fork and thence southeast toward the James. From the drainage pattern, altitudes, and valley shapes it is evident that the South Fork of Moormans River beheaded Sawmill Run just east of Jarman Gap and thereby shifted the Blue Ridge drainage divide a short distance westward.

The principal cause of piracy here and in other localities along the Blue Ridge is the shorter courses and steeper gradients of the streams on the southeast slope. Stream diversions of this kind are small features, but it may be that this process has been operating for a long time and that the

streams which now cross the Blue Ridge have cut through in a similar manner.

West of the Blue Ridge.—The headwaters of the Potomac River are vigorously encroaching on the drainage areas of the Youghiogheny and

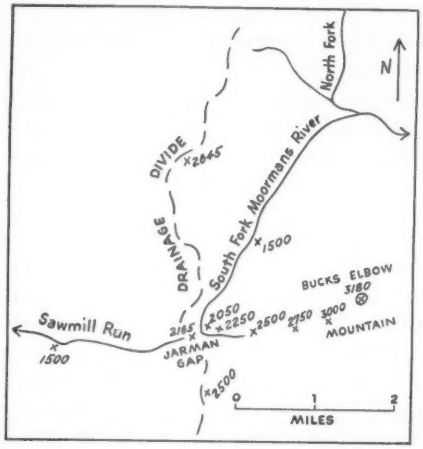

FIGURE 7.—*Recent piracy at Jarman Gap*
South Fork of Moormans River has beheaded Sawmill Run.

Cheat rivers in the eastern part of the Appalachian Plateaus. In the northwest corner of the Franklin, West Virginia-Virginia quadrangle, Seneca Creek, Roaring Creek, and other tributaries of the North Fork of the Potomac River are cutting back into the Allegheny Front and undercutting the heads of Red Creek, a tributary of Cheat River. The gradient from the divide down Seneca Creek is 1500 feet in 2 miles, whereas it is 700 feet in the same distance down Red Creek. On the Grantsville, Maryland-Pennsylvania quadrangle, the tributaries of Savage River (Potomac) are undercutting the tributaries of Castleman River (Youghiogheny) in Meadow Mountain. The Savage tributaries are flowing in deep, steep-sided, steep-gradient valleys, and the Castleman tributaries in shallower, broader, gentler gradient valleys. The fall down Poplar Lick Run toward the Potomac is 520 feet in a mile, and 360 feet down Little Laurel Creek toward the Youghiogheny in the same distance. In one place the divide between the northwest and southeast drainage is in a swamp.

In some places the highest summits are southeast of the present drainage divide, indicating that some streams have worked headward faster than

the retreat of the whole escarpment and isolated portions of the upland that formerly sloped northwestward. For example, Spruce Knob (4860 feet), the highest point in the Appalachian Plateaus, is east of the drainage divide. Tributaries of the North Fork of the Potomac River have worked entirely around this mountain and bring the drainage from the northwestern side. The distance to the sea from the drainage divide is now about seven times as far by way of the Monongahela, Ohio, and Mississippi rivers as it is by way of the Potomac, so that further northwest shifting is inevitable.

The headwaters of the James are opposed by tributaries of the New River in the competition for territory in the watershed. Here the divide is located in the western part of the Ridge and Valley Province. Throughout most of the area of competition the James is opposed by the Greenbrier, a long subsequent tributary of the New. From a point on the divide west of Monterey, Virginia, the distance to the sea along the James is about one-sixth that by way of the New, Kanawha, Ohio, and Mississippi rivers. At Covington, Virginia, the James (Jackson above Clifton Forge) River has an altitude of about 1300 feet; 20 miles to the west, at Lewisburg, West Virginia, the Greenbrier is 1800 feet above the sea. Both rivers appear to be fairly well graded, so that further downcutting will be slow. The opposing tributaries of these two rivers are vigorously gnawing into the main divide on Allegheny Mountain, which rises to altitudes of more than 3000 feet. With the advantage of steeper gradient decidedly in favor of the James headwaters it seems evident that the divide must migrate westward regardless of rock structure and resistance.

Recent and imminent piracy by the little headwaters of the James are suggested by the stream arrangements and the distribution of altitudes at a number of places along the divide. Along the Virginia-West Virginia boundary in the Monterey quadrangle, Warwick Run has probably beheaded Knapp Creek and is now undercutting Schock Run. From the divide the fall down Warwick Run is 1000 feet in less than a mile, whereas down Shock Run the fall is 1000 feet in 2½ miles. Other tributaries of Back Creek (Jackson drainage) have much steeper gradients than those of the Greenbrier. On the Lewisburg sheet it appears that the head of Back Creek has recently diverted the head of Little Devil Creek toward the James. The head of Hughes Creek has probably captured a tributary of Meadow Creek at Ruckers Gap.

The elbow of capture, often one of the striking evidences of recent capture, is difficult to recognize in this area of folded rocks. The rectangular pattern of streams is not disturbed by piracy. One must depend on such things as the slope of the land, the gradients of the streams, aban-

doned valleys and gaps, and former erosion surfaces, to recognize the piracy. On the basis of former erosion surfaces, Stose (1922, p. 23) points out that some of the valleys now drained by the headwaters of the James were formed when they were tributary to the New River and since capture their floors have been deeply dissected.

Not only is the James stealing from the drainage area of the Greenbrier, but the Greenbrier is robbing the Cheat River. The little lateral tributaries of the Greenbrier are vigorously cutting into the scarp of the Allegheny Front and undermining the high-lying Shavers Fork of the Cheat River. The most striking example of imminent capture here, where Leatherbark Run is about to divert the head of the Cheat River, has been described by Wright (1925, p. 55). When this piracy is consummated a beautiful elbow of capture will be produced.

The divide between the Roanoke and New rivers is in the Great Valley west of Roanoke, Virginia, and there is clear evidence that the New River has recently lost territory to the Roanoke.

Wright (1934, p. 61-70) describes in detail the post-Harrisburg diversion of about 200 square miles of territory from the New River to the Roanoke. He gives the shorter distance to the sea by way of the Roanoke and the Tuscarora quartzite barrier at Pearisburg, which holds up the New, as the principal factors in the capture. The fact that the grade of the Roanoke at Lafayette is about 500 feet lower than the grade of the New at Radford, 20 miles to the west, indicates that the westward-growing headwaters of the Roanoke may eventually capture the entire headwaters of the New above Radford.

Natural Tunnel, in Scott County, Virginia, represents an underground piracy (Wright, 1936, p. 129). Natural Bridge, in Rockbridge County, Virginia, is a remnant of the roof of an underground tunnel formed by sink piracy (Wright, 1934, p. 73-101). These features and other evidences of underground drainage in limestone regions suggest that subsurface piracy may have been locally important in shifting the main drainage divide westward.

Captures that can now be definitely recognized belong to comparatively recent geologic time. Evidences of more ancient captures have been obscured by erosion. On the basis of the present and recent tendencies and other evidences previously mentioned, it seems reasonable to infer that the Atlantic-Gulf drainage divide has slowly marched westward from its original position by progressive headwater piracy. The westward shifting of the divide may well have been intermittent, possibly at times reversed, depending on the help or hindrance of structure, rock resistance, and diastrophic movements.

CONCLUSION

The working hypothesis of Southern Appalachian drainage evolution here advanced includes the following principal parts:

(1) Location of the early Jurassic northwest-southeast drainage divide in the Blue Ridge anticlinorium, as a result of the Appalachian Revolution and Palisades Disturbance.

(2) Northwest migration of the divide by headward growth of the Atlantic-slope streams and progressive piracy of the northwest-flowing streams, because of the asymmetric position of the original divide and subsequent repeated eastward tilting of the Atlantic border region.

(3) The formation of many gaps by headward erosion.

(4) Local superposition of some streams by downcutting from a weaker formation.

This hypothesis is not entirely new. Some aspects agree with Davis' (1889) early interpretation, and others conform with recent proposals by Meyerhoff (1936); but it differs from both these views on some important points. As compared with the hypothesis of regional superposition, it is not dependent on the widespread perfect peneplanation or the widespread submergence of the Appalachian Highlands. In future studies of Appalachian drainage history it is hoped that the hypothesis here presented, as well as others, will be given due consideration.

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Hunter College, New York, N. Y.

Manuscript Received by the Secretary of the Society, July 6, 1938.